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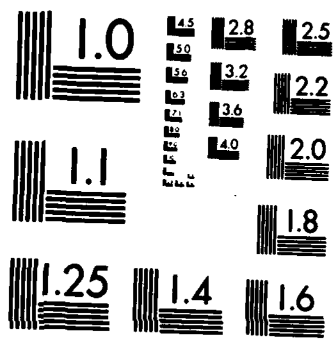

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OUTLOOK FOR IMPROVED NUMERICAL WEATHER PREDICTION USING SATELLITE DATA WITH A SPECIAL EMPHASIS ON THE HYDROLOGICAL VARIABLES

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This review critically surveys types and accuracies of available satellite data, satellite data retrieval methodologies, and data requirements of numerical weather prediction (NWP) models. NWP modelling, parameterization and data assimilation methods are reviewed with particular attention to the hydrological variables. It is concluded that the means exist to obtain improved retrievals of cloud and moisture parameters and to better parameterize the hydrological cycle. Further advances in the retrieval of humidity profiles, the parameterizations of clouds and the objective analysis of humidity are required.

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PREFACE

We have reviewed a substantial amount of the voluminous recent literature relevant to the assimilation and use of satellite data in updating numerical weather prediction models and are impressed with both the magnitude and the disconnectedness of the documented effort.

In particular, the tools are mostly available for using satellite measurements to obtain cloud amount and topography, but they have not yet been organized into a system for numerical weather prediction model calculations, much less for testing and tuning the model cloud parameterizations. Very little has been done on using satellite-derived surface characteristics, and it has not yet been clearly demonstrated whether and to what extent useful moisture retrievals can be obtained from radiances in present-day moisture channels.

Despite these shortcomings, the evidence that improved retrieval and assimilation of satellite-derived data and improved physical parameterizations will lead to better forecasts at all time scales is substantial, though not yet overwhelming.

The tasks of retrieving detailed atmospheric structure and constitution, including clouds and other water substance from satellite measurements, of assimilating the retrieved data into analyses, and of using them to test and tune the parameterization of clouds, boundary layer processes and surface conditions are challenging, but the prospect of significant success is encouraging. This review has, if anything, left us even more strongly committed to our proposed program, and made us aware of several promising avenues of research development. We hope it will also be useful to others engaged in improving numerical weather prediction.

We thank all of the people who have helped us to prepare this review. M. Cane and M. Chahine carefully reviewed the manuscript and provided us with insightful and expert comments. L. Bengsston, J. Coakley, J. Gyakum, M. Halem, A. Heymsfield, N. Phillips, W. Stern and J. Tribbia freely and considerately shared their recent pre-publication research results with us. M. Ghil helped R. Hoffman prepare an earlier version of Section 5. Andrew Metz performed the initial literature search for us with energy and dedication. Lisa Gibson carefully and patiently prepared the manuscript and figures. Finally we thank K. Hardy for support and encouragement while we prepared this review.

This review is dedicated to the memory of Andrew D. Metz.

ABSTRACT

Common large-scale numerical weather prediction (NWP) practice and research have relegated moist processes and variables to secondary roles. Now, with a prelude of three decades' experiences and steady advances in NWP, the greatest opportunity to advance global NWP probably lies in improving the parameterizations of moist processes and the observation and assimilation of moist variables. The more complete use of satellite data to test and tune the physical parameterizations as well as to provide more comprehensive initial states and boundary conditions promises greater understanding and better predictions of short-range weather changes.

Modern NWP models are often unable to predict rapidly developing synoptic scale intense storms. This defect is the most spectacular deficiency of current models and suggests some physical process or interaction between physical processes is missing from the models. The most likely cause is that latent heating is not properly parameterized in current models. Accurate prediction of latent heating is therefore of utmost importance for improving current global NWP models.

The prediction of clouds by modern NWP models is not better than a forecast of persistence out to 48 h. Cloud forecasts are of great intrinsic value - obviously so for aviation interests but also for forecasts of precipitation and minimum and maximum surface temperatures. There are two complementary approaches to improving cloud forecasts: improving the forecast of humidity and improving parameterization of cloud physics.

Thus, the parameterizations of all moist processes in global NWP models should be improved. The means exist: Some GCMs used in climate simulations and some research mesoscale models have advanced parameterizations which can be adapted for global NWP models.

Improved moisture analyses are needed. Satellite observations of moisture variables are essential. There are many observations of moisture potentially available from satellites which are not commonly used in global NWP models. Since satellite data is incomplete and asynchronous, four-dimensional assimilation methods must be used. Better statistics and initialization procedures for moisture variables are needed. Improved boundary condition data sets are also needed. These data are available or are potentially available from satellite observations. High spatial resolution is vital when moisture is involved because it varies more on smaller spatial scales than other meteorological variables.

Benefits that we expect to follow from the improvements discussed above include the following: (1) improved forecasts of intense storms; (2) better precipitation and cloud forecasts; (3) better initial and boundary conditions for mesoscale models; (4) better long-range forecasts.

1. INTRODUCTION

A three-year (8 man year) effort is currently underway at AER to enhance the capability of the AFGL spectral numerical weather prediction model (Brenner et al., 1982) to assimilate and use satellite data. This will be done by improving the manner in which satellite data are used to update the model and to interact with its physical parameterizations. Our goal is to test the full potential of satellite-based data for improving numerical weather prediction.

Algorithms will be optimized to retrieve temperature, moisture, surface properties and cloud amounts, heights and condensed water content from simultaneous combined infrared, microwave and visible radiances from DMSP and other satellites. Parameterization of radiative transfer, boundary layer processes, moist convection, clouds and moisture transport will be optimized to use the satellite data effectively, and tuned and tested by comparisons with initialized analyses, with forecasts, and with the data base enhanced by additional satellite observations.

A complete numerical weather prediction (NWP) system, such as the AFGL spectral system derived from the original NMC design (Sela, 1980), incorporates a great many individual elements, all of which must interact with each other properly and efficiently, especially in a test of the usefulness of improved satellite data assimilation. The forecast model integrates the equations of motion supplemented by the equations of state, hydrostatics and conservation of mass and water, the first law of thermodynamics, and parameterizations of physical processes that cannot otherwise be included explicitly. The role of water substance in such a model and in the real atmosphere is highlighted in Fig. 1.1. The data assimilation scheme updates the forecast with observations. When there is a variety of observing systems (e.g., radiosonde, aircraft, satellite) with different error characteristics, an optimum interpolation (OI) approach to data analysis is desirable. It is also particularly important that the physical parameterizations in the model be realistically sensitive to the satellite retrievals. Finally, the OI analysis should be initialized to be compatible with the model and prevent the growth of spurious gravity waves. This can be accomplished by applying a non-linear normal mode initialization to the analysis.

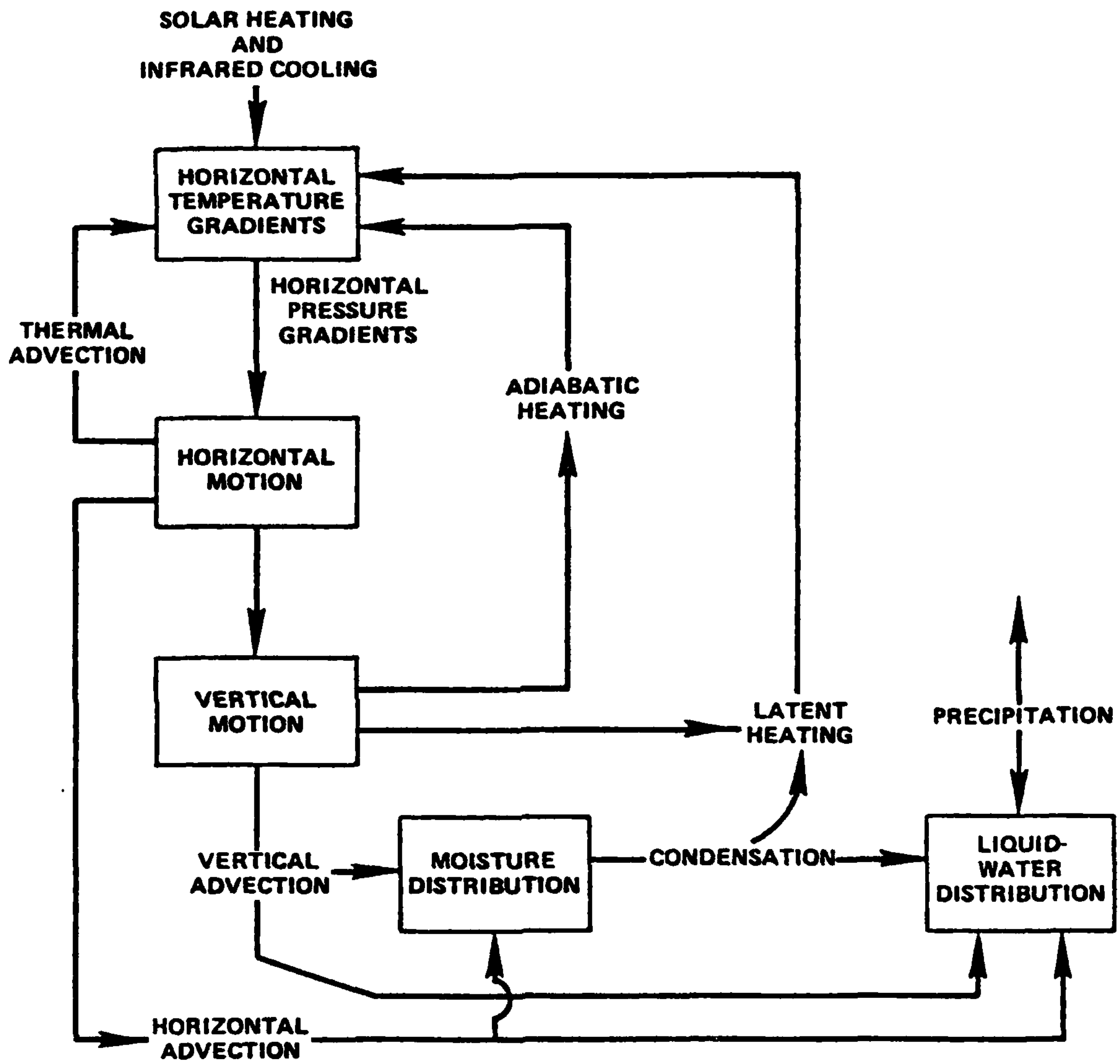

Fig. 1.1 The role of moisture and liquid water in a synoptic system (from Atlas and Thiele, 1981).

INTRODUCTION

In preparation for our planned research we have critically reviewed the available meteorological literature pertinent to our goals. It is clear from the above considerations that our future work touches on many aspects of the NWP problem. In this review we will cover all elements of a complete NWP system, but in view of our proposed research we will focus on those aspects of each element that involve the use of satellite data and/or involve the treatment of water substance in the atmosphere. It is evident that there are three areas of NWP which most concern us: retrieval of geophysical parameters, especially moisture parameters; parameterization of physical processes that involve these parameters; and assimilation of the observations into the model. In order to provide a context for our review we will now introduce the most important concepts and concerns from each of these areas.

Meteorological satellites routinely measure radiances in CO_2 and O_2 bands in the infrared and microwave spectrum, respectively, and these radiances are routinely used to obtain temperature soundings. Assimilation of the temperature retrievals is beginning to lead to improved forecasts, and this improvement should continue with the use of more accurate physical retrieval methods. The meteorological satellites also measure radiances in water vapor and window regions of the spectrum. These radiances, together with those in the temperature channels and with the high resolution infrared imagery, can be used to retrieve moisture, cloud amounts, cloud top heights, and surface conditions; but retrievals of these quantities are not yet routinely assimilated in NWP models or used in testing or tuning physical parameterizations.

In principle, all the physical parameterizations affect the model forecast. The effect of latent heat release appears to be the most important process for short-term predictions. The dynamical consequences of latent heat release are clearly seen in the study by Maddox et al. (1981) of the case of 0000 GMT 25 April 1975, in which an intense mesoscale convective complex developed over the central Mississippi Valley. Forecasts with and without moisture were made with Perkey's (1976) regional model. After 6 forecast hours, difference fields (moist forecast-dry forecast) clearly display the effect of latent heating (Fig. 1.2). The latent heat release due to the precipitation (a) results in a strong upward motion (b) at 500 mb. At 200 mb adiabatic cooling forced by convection and radiative cooling from cloud tops results in a region of lower temperature (c). This is also a region of higher

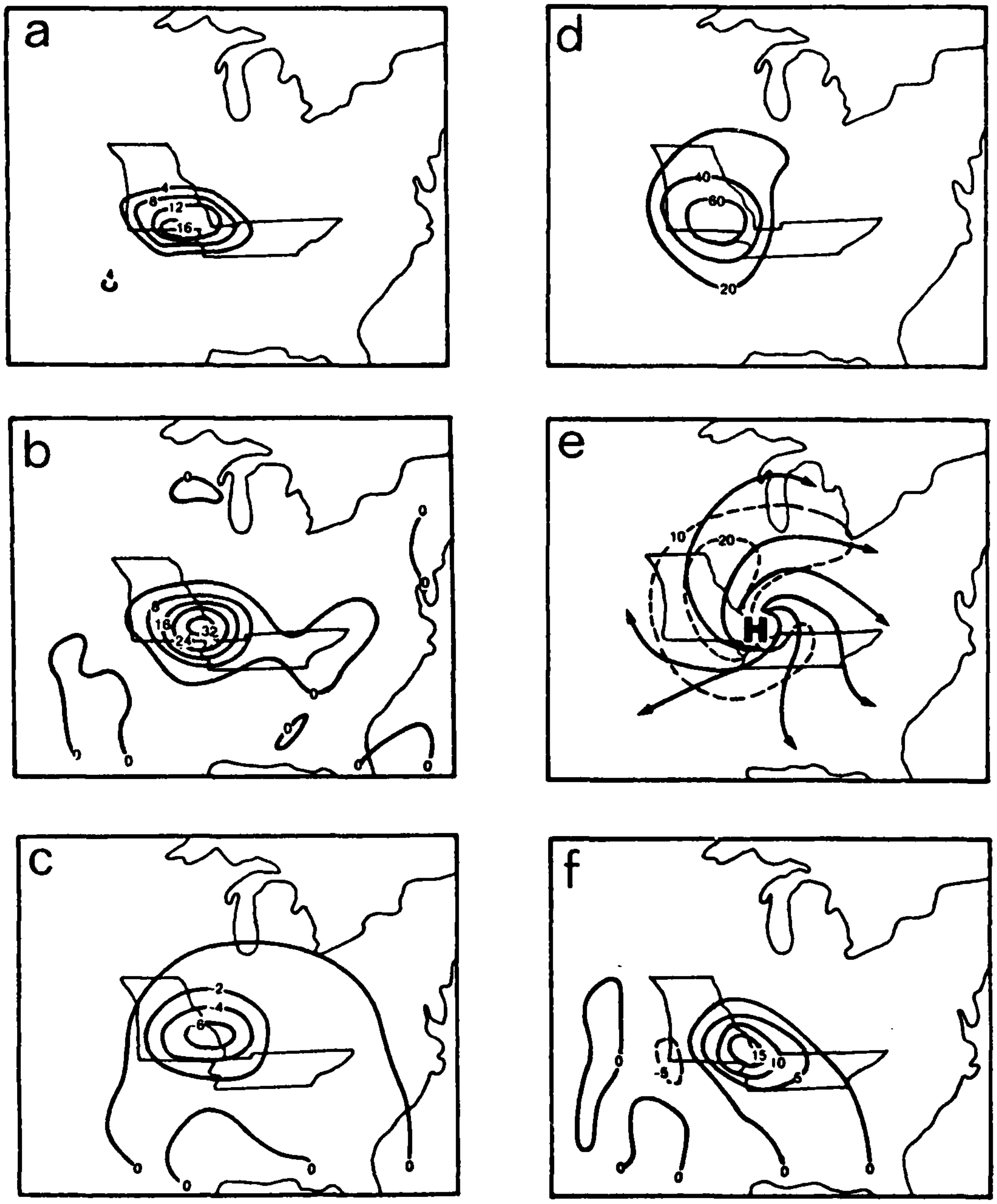

Fig. 1.2 Effect of convective latent heating. Differences (moist forecast - dry forecast) at 0600 GMT 25 April 1975 are shown for the (a) convective precipitation rate [mm $(10^4 s)^{-1}$], (b) 500 mb vertical velocity [cm s^{-1}], and the 200 mb (c) temperature [°C], (d) height [m], (e) winds [ms^{-1}] and (f) divergence [x 10^{-5} s^{-1}] (from Maddox et al., 1981).

geopotential heights (d) due to warming throughout the column. This high produces strong divergent outflow (e,f).

It is important to recognize that the model's physics also impacts the ability of an NWP system to assimilate data properly. Thus, the information contained in a satellite measurement will not be properly utilized if a model's physics does not correctly account for the process being observed. Optimum use of the satellite-based data of moisture, temperature, clouds, and surface characteristics in NWP models will depend critically on the capability of the model to incorporate this more complete data set so that a realistic link between the model and the data can be established. For this link to be made effectively, the radiation parameterization should be able to incorporate sub-grid scale fractional cloud amount, variable cloud altitude, cloud overlap and condensed water information derived from satellite data. In the parameterization of boundary layer processes, conservation equations for heat and water at the surface should be used to allow for the direct use of the satellite data-derived surface characteristics. Consistency between the clouds and hydrologic processes must also be maintained. Optimum use of satellite data will depend critically on the manner in which these processes are treated.

Parameterizations must be tuned. Simply stated, parameterization refers in general to the technique whereby the net effect of scales of behavior unresolved by a model upon those scales that are resolved is represented through use of simplified mathematical relationships between the scales. These simplified relationships typically involve arbitrary constants, whose values can be estimated on the basis of limited observations but which are often "tuned" within a particular model to yield the most reasonable looking results. As a result, there is considerable variety in these aspects of NWP models. Indeed, the amount of parameterized physics and the manner in which they are incorporated into a model can significantly impact the ability of the model to predict successfully those situations in which moist, viscid or diabatic processes are important. Some studies have been made to test the sensitivity of NWP models to the physics that is included and/or the parameterizations used, but much more along these lines needs to be done. In contrast, numerous sensitivity studies of this type have been performed in conjunction with general circulation models used to simulate climate fluctuations, which occur on time scales over which the physics are of paramount importance.

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It is vitally important to consider how the observations are put into the model. The problem of NWP may be stated as an initial value problem, but it is not practical to observe all the necessary variables simultaneously at all the relevant scales. Information present in previous observations must be used. Data assimilation is the process of obtaining a complete instantaneous three-dimensional state, i.e., description, of the atmosphere, from data distributed irregularly in space and time. Previously, data were observed synoptically so the distribution in time was regular and the problem was simply spatial. A solution to this simpler analysis problem, such as optimal interpolation, provides a part of the solution to the full four-dimensional problem. When asynoptic data, principally satellite data, are assimilated, it is necessary to use a numerical model to provide temporal continuity. In addition, using a model tends to make the model states obtained dynamically consistent or balanced. Whenever data are added, the model state is altered; this change is usually a shock to the model evolution because the observations inevitably contain some errors and because the model and nature, i.e., the real atmosphere, are not identical. This model imbalance results in two undesirable effects - the generation of gravity waves and the rejection of the data. In the initial value problem of NWP we may formally specify all the variables (e.g., surface pressure p_s, temperature T, and horizontal velocity $\underset{\sim}{V}$, for a dry atmosphere) with any values we like, but in reality, the atmosphere is in balance so that only some of the variables may be specified arbitrarily (e.g., p_s, T) and the others deduced diagnostically from the balance condition. The correct balance condition is known only for some very simple atmospheric models; however, even approximate balance is clearly desirable. A good approximation to balance is obtained by nonlinear normal mode initialization.

Interest in data assimilation really began with the study of Charney et al. (1969) who showed that if numerical weather prediction model runs are updated by model-produced temperature "measurements," the wind and pressure fields adjust towards their correct values, suggesting that a model can be initialized with satellite temperature soundings alone. This experiment has yet to be carried out with real data, but impact studies by GLAS have shown that updating by satellite and surface data and eliminating radiosonde data result in useful forecasts that can be comparable in accuracy to those using radiosonde and surface data but not satellite data (Halem et al., 1982). The

GLAS studies have also found small but positive impacts from adding satellite measurements in the data-sparse regions to available conventional data. This latter finding has been supported by recent studies (Jarroud et al., 1981) by ECMWF and by a marked improvement in ANMRC forecast skill scores after they started to use the southern hemisphere satellite and buoy data (Leslie et al., 1981).

In a broad outline we have attempted in our review to answer the following four questions:

1) What is the state of current operational global forecasting?
2) What data and/or physics are necessary to improve current forecasting capabilities?
3) To what extent can satellite observations provide these data?
4) How can these satellite data best be used?

The plan of this review is the following. In Section 2 we will discuss current operational and research NWP systems. We will also summarize the conventional data sources and their availability. In Section 3 we review the parameterization of moist physical processes and radiation in NWP models, and their impact on forecast skill. Section 4 reviews the retrieval of the geophysical parameters from satellite observing systems. The availability of the necessary data will be considered here. Finally, in Section 5 data assimilation and model initialization problems are reviewed. Section 6 contains a summary and conclusions.

2. CURRENT NWP PRACTICE

Predicting the weather is by no means an easy task: The atmosphere is a complex fluid medium whose evolution is governed by nonlinear dynamics, involving interactions between many scales and by myriad physical processes that occur at various scales. As a consequence, forecasts made by numerical weather prediction (NWP) models are very sensitive to the specification of the initial state and to the parameterizations of the physical processes. Errors in the initial state and the model's physics therefore lead to forecast errors. A third source of error is due to the combined effects of time and space discretization errors.

Forecast skill has increased steadily since the first operational forecasts, for quantities like 500 mb geopotential heights which characterize the large-scale flow patterns. Shuman (1978) documented this trend until 1976 (Fig. 2.1a). However, in recent years the skill of forecasts of precipitation occurrence has improved little (Shuman, 1978), if at all (Ramage, 1982), and the skill of quantitative precipitation forecasts has not improved (Fig. 2.1b).

Robert (1975) estimated that the error for short-range 500 mb geopotential height forecasts over North America could be apportioned as 48% due to discretization errors (38% horizontal, 9% vertical, 1% temporal), and 18% due to initial state errors. The remaining 34% of the total error Robert attributed to errors in the model's physics. These estimates were made for a 5-layer hemispheric primitive equation (PE) model with 381 km horizontal resolution. Since that time, no one has repeated Robert's study but some trends are clear: Discretization errors are certainly much reduced due to increased model resolution. Satellite data have decreased the initial state error; however, experiments with the FGGE data (Section 5.6) demonstrate that the impact of satellite data, at least as currently used, is not very large in the Northern hemisphere. Development of four-dimensional assimilation schemes and of physical parameterizations have undoubtedly reduced the initial state and physics errors, but the size of these reductions is unknown. Recently, it has become apparent that a significant portion of the forecast error is simply bias and that, surprisingly, much of this bias is at low wavenumbers. This bias grows with forecast time as the model drifts towards its own climate

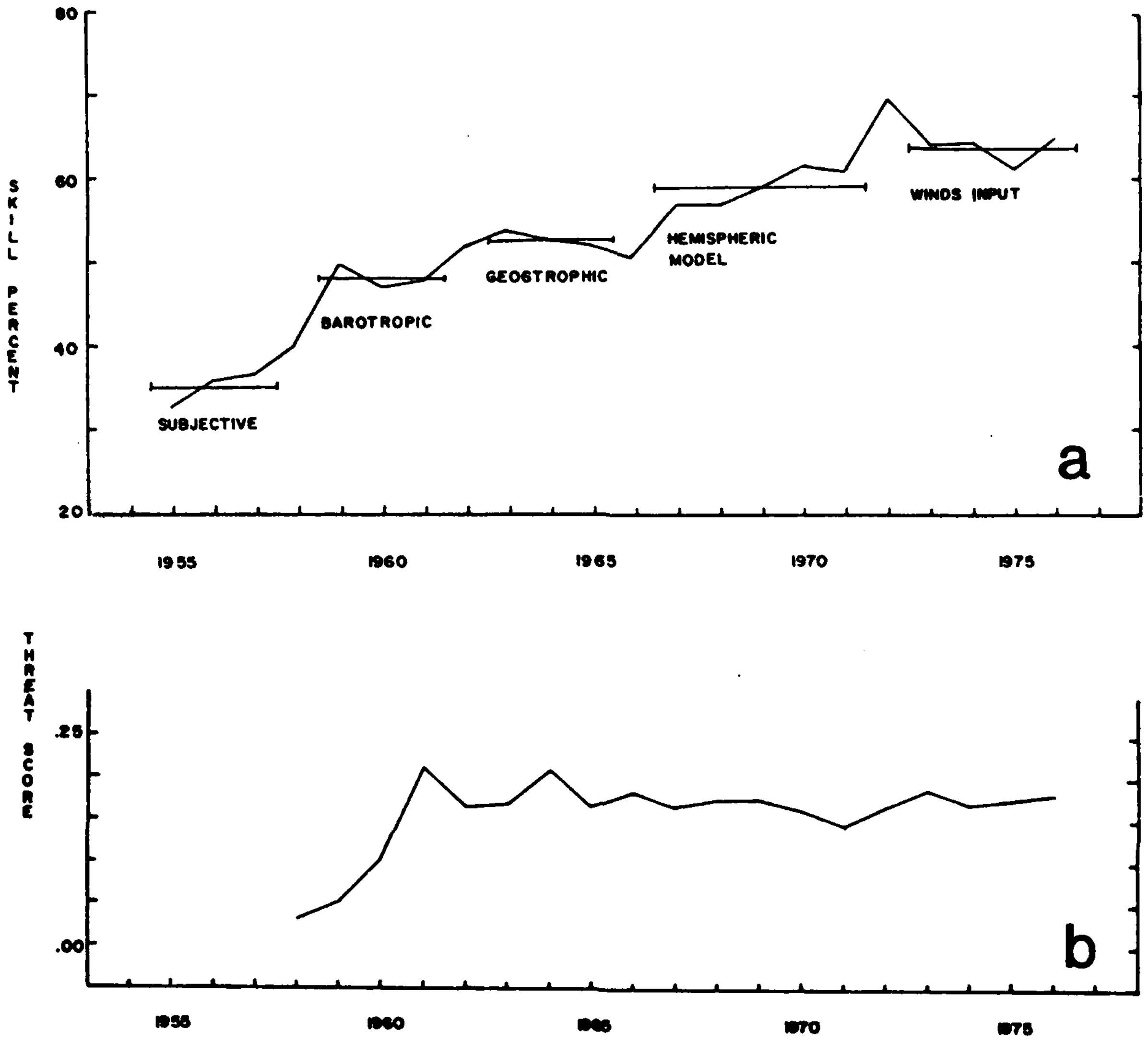

Fig. 2.1 Forecast skill (1955-1976). (a) Record of skill, averaged annually, of the NMC 36 h, 500 mb (~5.6 km high) circulation predictions over North America. The horizontal bars show averages for the years during which no major changes in models occurred. Prior to 1972 a quasi-geostrophic wind field was used for initial winds. Starting in 1972, analyses of observed winds were used instead. As shown, skill (percent) is 2 x $(70-S_1)$, which yields 0 for a worthless chart, and 100 for a virtually perfect one. (b) Record of skill of the NMC guidance for >25 mm of accumulated precipitation during the first 24 h after issuance. The measure of skill is threat score, which is the area of such precipitation correctly predicted, divided by (the total area less the area for which precipitation was neither observed nor predicted) (from Shuman, 1978).

(Simmons, 1982; Wallace and Woessner, 1981; Wallace et al., 1983; Arpe, 1983). This effect may be important for forecasts as short as 48 h. The model's climate, and hence this error, is determined principally by the model's physics.

Although often referred to simply as "models," current operational NWP schemes used to represent the global atmosphere and to predict its evolution are actually best thought of in terms of being complex systems, of which the forecast model is but one part. In addition to the forecast model, a complete NWP system includes procedures for (1) acquiring observational data from rawinsonde reports, satellites and other platforms, (2) checking these data for gross errors, (3) assimilating the data to produce an objective analysis of the state of the global atmosphere at a given moment in time, (4) ensuring the consistency of the objective analysis with the constraints imposed by the forecast model, and (5) producing output intended to diagnose the behavior of the NWP system. Although listed separately here, several of these procedures are actually closely linked in practice, as we shall discuss below while describing these parts of an operational NWP system.

The following discussion is not intended to be an extensive discourse on the theory of NWP; rather it will briefly review the basic elements involved in current NWP practice and will provide a context for the detailed discussions to follow. In subsequent sections we consider aspects of the model's physics and of the assimilation cycle in more detail. For more detailed discussions of computational aspects of NWP we refer the interested reader to the book edited by Chang (1977), to the textbook of Haltiner and Williams (1980) and to the recent review by Cullen (1983). In Chang's (1977) book we especially recommend the articles by Kasahara (1977) and by Arakawa and Lamb which describe grid point models, and the article by Bourke et al. (1977) which describes spectral models.

2.1 The Forecast Model

The heart of the NWP system is the forecast model which mathematically represents the fundamental physical laws governing the atmosphere. The model's primary role is to predict the future evolution of the atmosphere. In addition, the model serves as the vehicle for carrying data forward in time in an assimilation cycle. (More will be said of this second role shortly.) The model formulae consist of the primitive (i.e., unfiltered) equations of motion, along with equations expressing the laws of thermodynamics, hydrostatics, and the conservation of mass and moisture. In spectral models, for computational purposes, the equations of motion are generally reformulated with the vorticity and the divergence of the horizontal wind field as dependent variables.

Until recently, the operational practice had been to solve the model formulae in finite-difference form over a regular three-dimensional mesh of grid points. The distance between grid points was chosen to allow the scales of motion of primary interest to be adequately resolved explicitly, while keeping computational requirements within reason. This implies a horizontal grid-point separation distance on the order of 250 km in mid-latitudes, a value that also determines the size of the time step required to maintain the stability of the finite difference solutions. In the vertical, it is common to use normalized pressure (σ) as the vertical coordinate (Phillips, 1957) in order to simplify the mathematical boundary condition at the ground (Holton, 1979, p. 199). The most advanced operational grid point models typically utilized nine layers between the surface and 50 mb. Most of the σ surfaces were placed within the free troposphere, but their exact locations were often determined by the resolution deemed desirable in the boundary layer and stratosphere. At NMC, the tropopause was selected as a material surface in order to enhance vertical resolution without adding computational layers (McPherson et al., 1979). At the ECMWF, a "hybrid" vertical coordinate system is being introduced which resembles sigma coordinates near the ground, but reduces to pressure coordinates in the stratosphere, far away from the surface topography. The advantages of this mixed scheme are described by Simmons and Strüfing (1981).

Efficient transform methods developed in the last decade have now made spectral modeling approaches feasible (Bourke, 1972; Sela, 1980; Gordon and Stern, 1982). Without these transform methods, the evaluation of the nonlinear products in a spectral model is very time-consuming. Spectral forecast models have become (or are becoming) standard at many operational centers. The transform grid utilized in these models must contain enough (Gaussian) latitude and longitude points to avoid aliasing during the inverse transform process (Bourke, 1974). The number of these points is a function of the number of spherical harmonics chosen for the spectral expansion of the various horizontal fields. The spectral expansion is generally truncated at a wavenumber that yields a horizontal resolution equivalent to that of the older finite-difference models. For example, a grid point model with a 5° longitude grid increment represents wavenumbers 0-36, but only wavenumbers up to 12 or 15 are handled accurately, while a spectral model that includes wavenumbers 0-15 makes no error in representing or in differentiating these explicitly resolved scales. A spectral model therefore has fewer degrees of freedom to integrate forward in time and fewer grid points in the transform domain at which the effects of physical processes must be calculated, compared to a grid point model with equivalent resolution. This is one reason why spectral models are so efficient. The vertical representation in the new generation of forecast models remains the discrete level approach. The time integration scheme most often used is semi-implicit, which is easy to implement in spectral models and which allows large time steps in spite of the presence of gravity waves in the model. This is the second principal reason for the efficiency of spectral models.

The spectral forecast model has several other advantages over its predecessors besides its relative speed: the absence of pole problems, the absence of truncation errors when calculating horizontal derivatives, and the ease with which both gravity wave and very small scale noise can be filtered from the model solutions. Gravity wave noise is an unavoidable by-product of matching imperfect data with an imperfect model (see Section 5). The earlier operational finite-difference models were typically characterized by a high level of noise, and although several attempts were made to correct this problem, none proved entirely satisfactory (Kistler and Parrish, 1982). Modern spectral models, on the other hand, can easily make use of new normal mode initialization techniques to control noise. The normal mode techniques have been applied to both spectral and grid point models but their application to spectral models is more natural and easier since there are fewer modes to calculate.

It is in the area of the physics that current operational forecast models differ most; except for resolution, the dry inviscid adiabatic portions of the equations are treated in much the same manner by all the major centers. The term "physics" has a specialized meaning in the context of NWP. It refers to the representation of those atmospheric processes that serve as sources or sinks of momentum, heat or water vapor in the model equations. Such processes include those that contribute to the diabatic heating term in the thermodynamic equation (radiative processes, latent heating, transfer of sensible heat from the ground, etc.), to the friction term in the vorticity and divergence equations (viscous processes, "cumulus" friction, etc.), and to the moisture source/sink terms in the water vapor equation (evaporation, precipitation).

Among the difficulties involved in correctly incorporating physics into a model is the need to account for behavior on scales smaller than that resolved by the model's grid. For example, the exchange of momentum and heat between the surface and the atmosphere within the boundary layer results from a number of sub-grid scale phenomena. It is not feasible, of course, to account for each cumulus cloud explicitly within the context of global NWP; however, the cumulative effect of all the cumulus-scale convection on the larger scales of motion can be estimated through parameterization.

Parameterizations depend on model resolution; parameterizations must be tuned for each different resolution. Generally, models with high resolution have the most complete and accurate parameterizations. As model resolution increases more scales are explicitly represented so there is less to parameterize. However, with high resolution one obtains high accuracy in the dynamics and generally one attempts to obtain comparable accuracy in the physics.

Why are these physical processes treated so differently in the various models? Fundamentally, we do not know the correct way to formulate the mathematics that describes the moist, viscid, diabatic physical processes relevant to the NWP problem. For example, within the context of a global model it is not at all obvious how much rain should evaporate as it falls. Most parameterizations assume a separation of time and horizontal space scales, i.e., the physical processes being parameterized are assumed to equilibrate so fast and to act on such small horizontal scales that they can be estimated from the explicit model variables in a single vertical column at a single time. This assumption is not always appropriate, but the alternative is to explicitly model the physical process. It is equally true that we do not completely understand which of the myriad physical processes operating in the atmosphere are important for weather forecasting. Thus, for example, some would argue that specifying the correct emissivity of cirrus clouds will improve a model's performance, but most would agree that incorporating storm electrification would not. Also, there is the practical consideration of limited computer resource available to an operational center, because a model's "physics" typically requires a large amount of computer time.

2.2 Data Sources and Processing

There are two types of data pertinent to an NWP effort. The first consists of those "real-time" observations used to update the analysis of the current state of the atmosphere. The second consists of those more slowly varying fields that are needed to define the "boundary" conditions (e.g., sea surface temperature) within the model's physics. A number of simulation impact experiments were made prior to the First GARP Global Experiment (FGGE) to identify the accuracies required of both data types for NWP in this decade. Once FGGE was completed, real data impact experiments were conducted (Section 5.6). In Table 2.1 (adopted from National Academy of Sciences, 1983), we list current estimates of the observational accuracies desired for operational NWP in the next ten years. Not all of these requirements are currently met on a global basis, although they are potentially within the grasp of the satellite.

The collection and dissemination of routine real-time observations have become, under the auspices of the WMO, highly systemized efforts. Data from rawinsondes, pilot balloons, surface stations, buoys, ships, aircraft and satellites are collected regularly as part of the Global Observing System of the WMO's World Weather Watch (WWW). These data are then transmitted over the WWW's Global Telecommunication System (GTS) to the various national and regional operational weather forecast centers. In theory, therefore, all of the world's forecast centers have the same basic data available to them. In practice, however, the capacity of the GTS is limited so that not all of the satellite data are transmitted. Hence, centers like the U.S. NMC that acquire certain satellite data directly have access to higher resolution observations. Similarly, some station observations that are not reported within the GTS cutoff time limits may be reported directly to certain local meteorological centers.

Once received at an operational center, the real-time observations are fed into that center's own data handling streams. An early stage of this data handling includes checks for gross errors in data reports (Baker et al., 1981; Bengtsson et al., 1982a). Such checks include verifying the location of sta-

Table 2.1 Desired observational requirements. The desired resolution and accuracy are given for measurements in the troposphere for climate, large-scale dynamical studies and weather prediction (from National Academy of Sciences, 1983).

Parameter	Temporal Resolution	Horizontal Resolution	Vertical Resolution	Accuracy
Horizontal velocity $\underset{\sim}{V}$	12 h	100 km	1 km	$\pm$ 3 m s^{-1}
Temperature T	12 h	100 km	1 km	$\pm$ 1° C
Specific humidity q	12 h	100 km	1 km	$\pm$ 1 g kg^{-1} or $\pm$ 10%
Relative humidity	24 h	100 km	1 km	$\pm$ 10%
Surface pressure p_s	12 h	100 km	-	$\pm$ 1 mb
Sea surface temperature	5 days	100 km	-	$\pm$ 1° C
Land surface temperature	12 h	100 km	-	$\pm$ 2° C
Surface moisture availability (% of saturation)	24 h	100 km	-	$\pm$ 10%
Clouds (% coverage)	3-6 h	100 km	500 m	$\pm$ 5%
Top of atmosphere radiation budget	3-6 h	100 km	-	$\pm$ 5 W m^{-2}
Surface heat and moisture fluxes	12 h	100 km	-	$\pm$ 10 W m^{-2}
Precipitation (total)	1 day	100 km 25 km	-	$\pm$ 5% oceans $\pm$ 10% land

tions (for example, ensuring that ship positions are over water), testing that temperature soundings satisfy hydrostatic constraints, and "flagging" values that fall outside reasonable bounds. Other error checks on the data, such as comparisons among neighboring values ("buddy checks"), are generally performed later within the analysis.

Various types of auxiliary meteorological and geographical data fields are necessary for specifying conditions at the boundaries of an NWP domain. In the global sense, it is important either to fix surface conditions a priori or to generate them in the course of the NWP modeling. Unlike the situation for the real-time observations, the data required for the model's physics are far from being routinely available. (Indeed, one reason for our lack of understanding of much of these physics is the paucity of relevant data sets.) By and large, data sets of such quantitites as sea-surface temperature (SST), surface albedo, sea-ice extent, and soil and vegetation type are derived from independent observations or climatological records. Often the decision as to what auxiliary data sets will be utilized is based on the sophistication of the individual model's physics. If soil moisture, for example, is treated very crudely by a model, then there is little need to do more than simply distinguish land surfaces from ocean. On the other hand, if soil moisture is handled in some detail, then a global field of soil types is desirable. Sea surface temperatures may be chosen from a climatological record or may instead be updated by near real-time observations. Similarly, clouds may be based on climatology or, as is becoming the state-of-the-art, may be predicted by the model equations themselves. Data regarding snow, sea-ice extent and land surface albedo are also derived from different sources by each center. It is fair to say, in general, that the situation with regard to the auxiliary data is constantly evolving, particularly as the operational centers seek new ways to improve the physics in their models.

SST climatologies on spatial scales of at least 500 km are reviewed by Reynolds (1983). He summarizes the available data sets from oceanographic stations and bathythermographs (cast data) and (less accurate) surface marine weather data, as well as satellite-derived SST fields. Levitus' (1982) 1° latitude-longitude grids are derived from cast data in the National Oceanographic Data Center; Robinson (1976) and Robinson et al. (1979) add shore station data and surface marine data, and Reynolds (1982) uses both data types from available data sets at the National Climatic Center (NCC). Other climatologies have more extensive surface marine observations (including Alexander and Mobley (1976) and Rasmusson and Carpenter's (1982) tropical Pacific strip).

SST data are also computed from satellite observations (see Section 4.2). While the coverage is much fuller than that from ships, the accuracy is much less (up to 4°C), and the length of record since the appearance of the first satellite derived SSTs is rather short for production of a climatology. Although SSTs from satellites are not available in regions of cloud cover, they are still much more extensive than those from ships, which are largely confined to parts of the North Atlantic and North Pacific and the major shipping lanes (Miyakoda and Rosati, 1982).

The albedo of ground and ocean surfaces is used in the calculation of radiative heating in the atmosphere and in the calculation of ground temperature. Estimates of zonally averaged monthly mean surface albedos (averaged over the full spectrum of incoming solar radiation) were prepared by Kukla and Robinson (1980), and these are stratified by 2° latitude bands. Values range from a low of near 6% in the summertime Southern Ocean latitudes to a high of 80% over the fully ice-covered polar regions. The global mean values are 17.5 and 16.1% for January and July, respectively. Robock's (1980) calculations of surface albedo are energy weighted over the spectrum. Briegleb and Ramanathan (1982) developed a surface albedo model within the context of their planetary albedo calculations. These results are based on a two-dimensional global surface distribution.

All the above computations of surface albedo depend critically on estimates of the specific local characteristics of the Earth's surface. Thus, ice-free ocean, sea-ice, snow cover over land, and snow-free land of different vegetation and soil characteristics all have different albedos which, in turn, vary with wavelength. Briegleb and Ramanathan (1982) cite the distribution and various albedo values of soil and vegetation types, with a number of references. Sea-ice albedo characteristics are summarized by Grenfell and Maykut (1977).

Vegetation and land use types for use in a variety of studies have been compiled on a 1° latitude-longitude grid by Matthews (1983) and are available through the NCAR archives. In this archive, vegetation and land use in an area fall in one of the 148 categories of the UNESCO (1973) classification scheme. An earlier digital compilation of vegetation produced by Hummel and Keck (1979) was specially designed for use in albedo studies.

Sea-ice distributions are analyzed on weekly charts by the Navy-NOAA Joint Ice Center (Godin and Barnett, 1979). Sea-ice grids have also been described by Walsh and Johnson (1979) for the Arctic. Variability in the Antarctic is discussed by Ropelewski (1983) and in the North Pacific and North Atlantic by Walsh and Sater (1981).

Finally, topography must be specified. Small scales in the Earth's topography affect the atmosphere and the effect of these scales should be parameterized. Until recently, most model topographies have been filtered versions (e.g., Pitcher et al., 1983) of the Scripps' 1° x 1° topography (Smith et al., 1966). Now the U.S. Navy's 10' x 10' topography is available. Recent experiments at ECMWF have demonstrated that the climatology of the model is sensitive to the representation of topography used and that an improved "envelope" topography has a positive impact on forecast skill (Wallace et al., 1983). This envelope topography is calculated from the Navy topography as the sum of the local mean (i.e., grid box averaged) topography and twice the local standard deviation of topography. This topography is supposed to account for the fact that air sheltered in valleys does not interact much with the large scale flow.

2.3 Assimilation and Initialization

Numerical weather prediction is primarily an initial value problem; i.e., given the present state of the atmosphere, the object is to solve the prognostic system of equations for a future state. It is in this context that the real-time data discussed in Section 2.2 must be blended with the model described in Section 2.1. This blending is accomplished through the steps of analysis and initialization, within an assimilation cycle.

A priori, it would appear possible to treat the objective data analysis task separately and independently from that of the numerical forecast. In modern practice, however, the two tasks are intimately entwined in an assimilation cycle. In particular, the fields generated by the forecast model serve as first guesses for the objective data analysis. Clearly, this approach provides a convenient means for filling gaps in data sparse areas. It also offers a convenient framework for handling asynoptic data, since the forecast model can be readily updated with such information. Most importantly, however, modern data assimilation is designed to incorporate atmospheric observations from a wide variety of instruments into the model-generated fields without creating excessive noise in the model's forecast of those fields.

Noise in a forecast model can arise in a number of ways. Because gravity waves represent solutions to the primitive equations used by operational models, imbalances between the mass and motion fields may excite these waves and lead to their propagation through the forecast cycle. Approximations to the balance observed in the atmosphere are the familiar geostropic and gradient wind balances. These relationships between mass and momentum are expressions of the approximate balance between the Coriolis force and the pressure gradient force, and, in the case of gradient wind balance, the centrifugal force in the atmosphere. In some situations these forces do not even remotely balance each other because the effects of physical processes must be considered. For example, friction in the boundary layer and cumulus latent heating in the Intertropical Convergence Zone are important factors in the balance of the atmosphere. Inasmuch as gravity waves do not carry any information of meteorological importance, they may be regarded as noise interfering with the clear resolution of the meteorologically significant waves. Imbalances between the mass and motion fields in the model are introduced whenever real data are inserted that alter portions of these fields. A central goal of the data assimilation process, therefore, is to minimize the degree of imbalance "seen" by the forecast model. Because the physical processes are involved in the balance of forces, especially in the tropics, accurate parameterizations are needed to achieve this goal.

Another source of noise and error in the data assimilation process is due to the different vertical coordinates used by the forecast model and by the global observing network. As we discussed earlier, the forecast model uses a σ coordinate system. Data, however, are routinely reported and analyzed in pressure coordinates. Techniques to interpolate (or extrapolate) between σ and p surfaces (utilizing cubic splines, for example) have been developed and are continuing to be improved, but inevitably some errors are introduced by this process (Baker, 1983).

As we noted earlier, concurrent with the advent of spectral models, advances have been made in the battle to minimize noise. These techniques, which are classed under the category of normal mode initialization, take advantage of the very different propagation characteristics of gravity and meteorological waves. In general terms the approach involves expanding the fields produced by the data assimilation step in terms of the model equations' normal modes. Those modes associated with the fast propagating gravity waves can then be identified and removed before continuing the assimilation. Much current research is being devoted to finding the best means for accomplishing this filtering. Further details about data assimilation and initialization can be found in Section 5.

We have tacitly assumed in our discussion that observational data are being assimilated in an objective manner, with a minimal amount of manual intervention. Although this is generally the state-of-the-art at operational centers, there is at least one major center that is still relying on manual analyses of data, namely the Australian Bureau of Meteorology's National Meteorological Analysis Centre. Research at ANMRC into means for objectively assimilating Southern Hemisphere data as a routine approach does appear promising (Bourke et al., 1982), but for the moment the gaps in the Southern Hemisphere data coverage pose a considerable obstacle to operational objective analysis. Indeed, Zillman (1983) reports that subjective viewing of satellite imagery is invaluable to the Australian analyses for identifying the location of major fronts and depressions over the Southern Hemisphere. This, then, is a clear example of the importance of satellite cloud information to NWP.

2.4 Post-processing and Diagnostics

The output from the forecast model consists of global fields of the prognostic model variables on constant pressure surfaces. To translate this information into predictions of surface weather elements of interest to the public requires an additional step that has yet to be perfected. Indeed, it is at this stage that a great deal of subjectivity enters the system, giving rise to the concept of weather forecasting as an "art." Some objective approaches have been developed, such as the MOS (model output statistics) specifications used at NMC, in which surface weather elements are related to the forecast model output through statistical regressions based on historical data (Klein, 1982).

A great many diagnostic tools have been developed over the years to assess the performance of NWP systems. While subjective evaluations are potentially very useful, most published accounts focus on quantitative evaluations. Anthes (1983) has reviewed many of the quantitative measures used to evaluate forecasts in terms of skill and realism from the perspective of the regional modeler. His review is comprehensive and, for the most part, applicable to any atmospheric forecast model. Commonly used quantitative measures of forecast skill include S_1 scores, rms errors, correlation coefficients of forecast and observed changes, errors in predicting characteristic features (e.g., minimum pressure of a storm) and threat scores. Before these measures are calculated, the differences between the forecast and observed fields may be filtered spatially or temporally to focus on some particular scale. For reference, the skill scores obtained are usually compared to scores obtained by using a "forecast" of climatology or persistence. The most popular of these scores is the S_1 score (Teweles and Wobus, 1954) which is defined as

$$S_1 = 100 \sum |e_g| / \sum |G_L|$$

where the sums are over all pairs of adjacent verification points, e_g, is the error in the difference between adjacent points and G_L is the observed or forecast difference, whichever is larger, between adjacent points, for pressure or geopotential. S_1 is a measure of how poorly gradients of pressure or geopotential are predicted. In the extratropics S_1 is therefore a measure of how poorly winds are predicted. Because of its definition, no statistical theory of S_1 scores has been developed. The theory of rms and correlation measures of skill is discussed by Leith (1982). Murphy (1970, 1971) discusses ways to evaluate probability forecasts and Gordon (1982) describes a skill score applicable to any categorical forecast. Quantitative measures of forecast realism include comparing observed and forecast statistics, including autocorrelations, spectra (e.g., kinetic energy) and terms in budget calculations, and calculating skill scores for the forecast offset horizontally or temporally by different amounts in order to separate phase and amplitude errors. It is on the basis of these sorts of skill scores that reference is often made (e.g., Fig. 2.1) to the progress of NWP over the last decade.

Various diagnostic calculations are imbedded in the NWP system and can be used to monitor the performance of the assimilation/initialization routines. For example, vertical motion fields can be derived from the analyzed wind fields and then compared with synoptic maps of precipitation. The ECMWF has found that calculating rms-differences between the observations and the first guess analysis and initialized fields to be a powerful tool in examining the quality of their products over different areas (Bengtsson et al., 1982a). Finally, diagnostic calculations are important elements of the physics packages of NWP systems, allowing insights into approaches that will improve parameterizations.

3. MODEL PHYSICAL PARAMETERIZATIONS INVOLVING SATELLITE DATA

A complete numerical weather prediction model should include detailed and realistic parameterizations of physical processes, such as radiative transfer, boundary layer processes, moist convection, clouds and moisture transport, since it is the combined effects of these processes that drive the wind and weather systems by maintaining and altering the fields of available potential energy and potential vorticity. To the extent that these processes are important for short-range predictions, the cloud processes can be singled out as being particularly critical because they can greatly enhance the strength and localization of diabatic temperature changes, and greatly change the geographical distribution of net radiative flux at the top of the atmosphere.

Data from meteorological satellites confirm that clouds strongly influence the solar and thermal radiation and are organized by the large-scale flow. Because of the strong interactions between radiation, the circulation of the atmosphere, and clouds (see Fig. 3.1), it is important to investigate and improve the parameterization of these processes and their relationships.

Presently, even in the most comprehensive GCMs, these relationships have been crudely approximated due primarily to inadequate understanding of the physical processes which control cloud dynamics, the hydrological cycle and cloud-radiation interactions (cf. WMO, 1975, 1978; NASA, 1981). In particular, two closely related types of parameterizations (see Fig. 3.1) are needed in GCMs. The first is the cloud physical development and the second the cloud radiation field. A survey of the methods used in the NWP and climate models for determining amounts and properties of clouds has recently been conducted by WGNE (the WMO Joint Scientific Committee's Working Group on Numerical Experimentation). The resulting report, Studies of Cloud/Radiation Interaction in Numerical Models (WGNE, 1983) provides a detailed description of the cloud prediction scheme and radiative properties currently used in a dozen or more large scale numerical models around the world. Physical parameterizations for use in regional models are discussed by Anthes (1983).

In this section, we present a brief review of the physical parameterizations related to the moisture field and discuss their potential improvement through comparison of such quantities as cloud topography, horizontal coverage, phase, thermal emissivity and solar albedo with their retrievals from

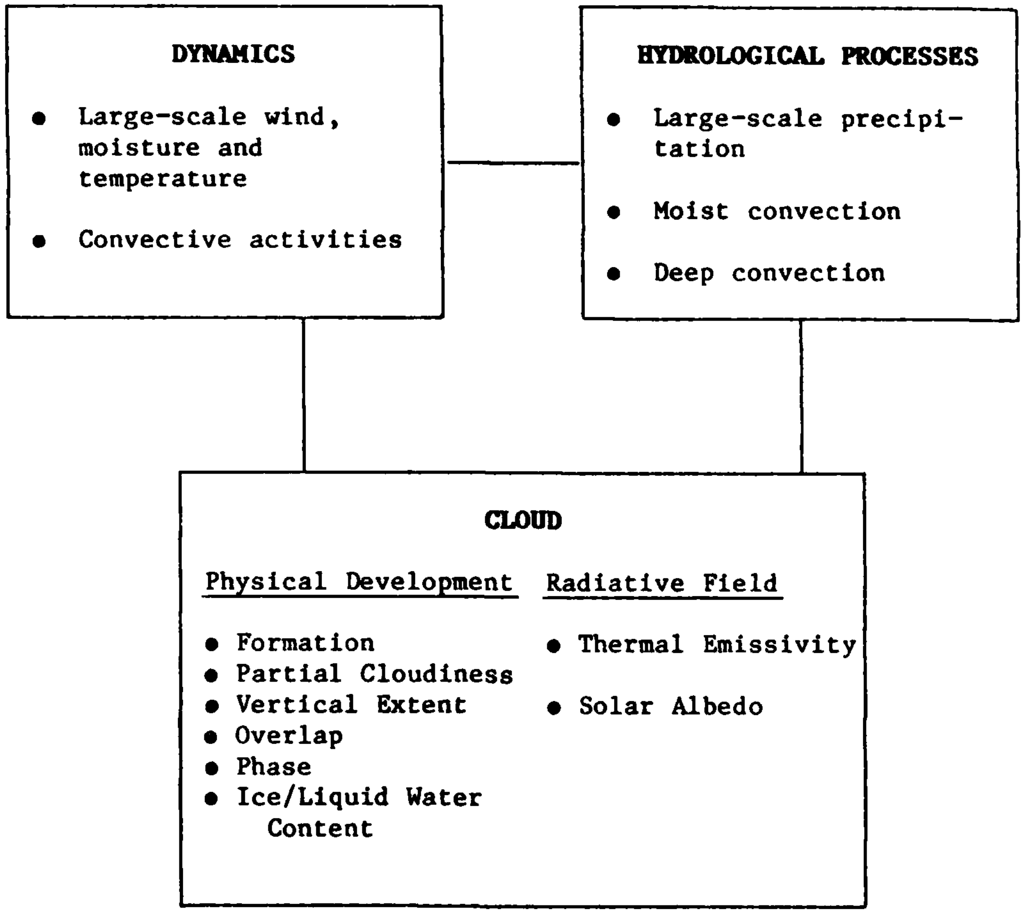

Fig. 3.1 Atmospheric processes. There are interactions between radiation, cloud and large-scale atmospheric variables.

satellite data, as described in Section 4. The emphasis in this section is on physical processes involving clouds for reasons given above, and because clear sky radiation and, to a large extent, clear sky boundary layer processes are better understood and are more easily parameterized special cases. A list of suggestions for the improvement of model physical parameterizations involving satellite data is provided at the end of this section.

3.1 Cloud Physical Development

According to their observed characteristics, clouds are generally grouped into four main types: (1) boundary layer clouds, (2) penetrating convective clouds, (3) clouds associated with cyclonic systems, and (4) upper level cloud systems. The ultimate goal is to model these cloud systems physically from generation through decay. Current GCMs all contain clouds explicitly, either specified entirely from climatology (and thus invariant), or derived from model predicted parameters, e.g., from relative humidity (RH) as first proposed by Smagorinsky (1960). Satellite data, such as distribution of moisture, cloudiness, temperature, etc., together with conventional data may be used to test and initialize these parameterizations, thus improving their realism.

3.1.1 Cloud Formation, Maintenance and Dissipation

In general, existing parameterizations assume that cloud will form if (1) RH exceeds some critical value RH_c and/or (2) the level of vertical convective activity exceeds a certain index. Some models employ only the first criterion (cf. WMO, 1978; WGNE, 1983). A brief review of the studies of these parameterizations is given here.

Marine Boundary Layer Clouds

Studies of this type of clouds are numerous, ranging from the simplest approach of Lilly's (1968) cloud-topped mixed layer model to the multi-layered model of Albrecht et al. (1979), to the more complex models of Deardorff (1980), Beniston and Sommeria (1981) and others. Herman and Goody (1976) have studied the formation and persistence of summertime Arctic stratus cloud.

These parameterizations for stratus (and cumulus) development include the sea surface fluxes of moisture and heat, advection, stability, radiative cooling and entrainment of air from above the cloud. Investigations of these mechanisms and of their relationship to fractional cloudiness remain important areas of research (cf. Suarez et al., 1983). Satellite data may play a key role in improving the parameterization of boundary layer stratocumulus clouds; high resolution infrared imagery measurements and procedures for intepretating these measurements, such as that of Coakley and Bretherton (1982), may be used over oceans to test and, if necessary, provide initial data for these parameterizations.

Deep Cumulus Clouds

It is known that cumulus convection redistributes sensible and latent heat and momentum in the vertical and is particularly important in the tropics. The parameterization has to relate the convective transports of heat, moisture and momentum by the cumulus clouds, which cannot be explicitly resolved by the large-scale model, to the variables predicted by the model.

Two classes of cumulus parameterization are employed: (1) convective adjustment and (2) cloud-model schemes. In the convective adjustment, the effect of cumulus convection on the environment is considered simply by adjusting the model's moisture and thermodynamic fields, whenever instability is detected, towards a moist neutral state in which a constant equivalent potential temperature is maintained (Manabe and Strickler, 1964; Kurihara, 1973). All the water condensed is assumed to fall out instantaneously. In the cloud model schemes, the effect of subgrid scale penetrative convection on the large-scale equations is considered. Once the distribution of cloud temperature, moisture and momentum is determined, their effect on the environment may be determined (Kuo, 1965, 1974; Ooyama, 1971; Ogura and Cho, 1973; Arakawa and Schubert, 1974; Molinari, 1981b). Anthes (1977) has presented a review of cumulus parameterization and proposed a one-dimensional cloud model to study the cloud-environment interactions. Recent studies of cumulus parameterization have emphasized the development of more physically realistic schemes (Geleyn et al., 1982; Giraud and Joly, 1982) and model sensitivity experiments (Helfand, 1979, 1981; Bennetts and Rawlins, 1981).

Middle and Upper Level Cloud Systems

Presently, these clouds are generated in GCMs as a result of grid-scale saturation, or as a result of sub-grid-scale convective processes. Because of the complex nature of clouds themselves and of their dependence on all other hydrologic processes including evaporation, condensation and precipitation, no satisfactory theory exists for relating the cloud distribution to large-scale wind, temperature and humidity. However, several observational studies are of interest, especially those of high cirrus clouds.

In the tropics high cloud forms initially in association with convective systems. Cumulonimbus clouds transfer water from low to high levels where they form spreading cirrus shields (cf. Houze, 1983). These high clouds appear to dissolve slowly and can exist for a long time after the dissipation of the cumulonimbus.

Observational studies (Orville and Hubbard, 1973; Heymsfield, 1975, 1977; Hallett et al., 1978; Ohtake et al., 1982; Heymsfield and Platt, 1983) reveal that the ice water content, ice crystal concentration and size distribution together with temperature and vertical air velocity could provide the basis for parameterization of cirrus cloud if the detailed balance of evaporation, condensation and precipitation is considered. In addition, Paltridge and Platt (1976) have indicated that the radiative process is an important factor in cirrus development after the onset of nucleation. Since much of this information (including ice/liquid water content, phase, horizontal coverage) is directly or indirectly available from satellite data, it should be possible to use this information for initial conditions and for testing to improve parameterizations for cirrus cloud development in GCMs.

Slingo (1980) has employed the GATE data to develop a cloud parameterization scheme for the UKMO 11-layer tropical model. The scheme allows for low, medium, high and convective clouds. The convective cloud has been related to the convective mass flux. The cloud fields produced by the model during a three-day forecast from real initial data showed many of the large-scale features of the observed cloud field. In particular, the scheme has successfully predicted the persistent stratocumulus fields occurring under the trade-wind inversion. Bennetts and Rawlins (1981) have parameterized mid-latitude cumulonimbus convection and examined its influence on the simulation of cloud development. Sundqvist (1978) developed a sophisticated parameterization scheme for non-convective condensation which includes cloud water content as a prognostic variable. The influence of such a scheme on model prediction has also been examined (see Sundqvist (1981) and Section 3.6).

3.1.2 Partial Cloudiness

Based on observational records, Smagorinsky (1960) has performed a probabilistic estimate of percent cloud cover as a linear function of relative humidity. Since then, models that predict fractional cloud cover, F, generally assume that

$$F = \begin{cases} 0 & RH < RH_c \\ f(RH) & RH > RH_c \end{cases} \qquad (3\text{-}1)$$

where f(RH) is a monotonically increasing function of relative humidity, RH. Albrecht (1981) has used such a relationship to predict partial coverage for marine boundary layer clouds. However, the results were found to be very sensitive to small change in RH values. A relationship of this form, with f quadratic in RH, has also been used by Slingo (1980) in the UKMO model, by Geleyn (1980) in the ECMWF model and by others (cf. WGNE, 1983). Relationships like (3-1) have been successfully used in simulating large scale cloud patterns (Fig. 3.2).

As an alternative to (3-1), an empirical relationship between fractional cloud cover and a moisture variable called condensation pressure spread (CPS) has been derived at AFGWC (cf. Mitchell and Warburton, 1983) based on surface data. CPS is defined by

$$CPS = p - p_s, \; p > p_s \qquad (3\text{-}2)$$

where p is the observed pressure and p_s is the lowest condensation level of a parcel of air with the observed pressure, temperature and dewpoint. CPS is larger for dry air and zero for saturated air. The empirical relationship between CPS and cloud amount shown in Fig. 3.3 is currently used in the AFGWC 5-LAYER forecasting model (Fye, 1978; Mitchell and Warburton, 1983).

3.1.3 Cloud Vertical Extent

The treatment of cloud in the vertical direction is usually either to assume the cloud to be infinitely thin or to occupy the complete model layer.

Recent studies of finite clouds (Harshvardhan, 1982; Ellingson, 1982; Harshvardhan and Weinman, 1982) have shown that the finite cloud radiation

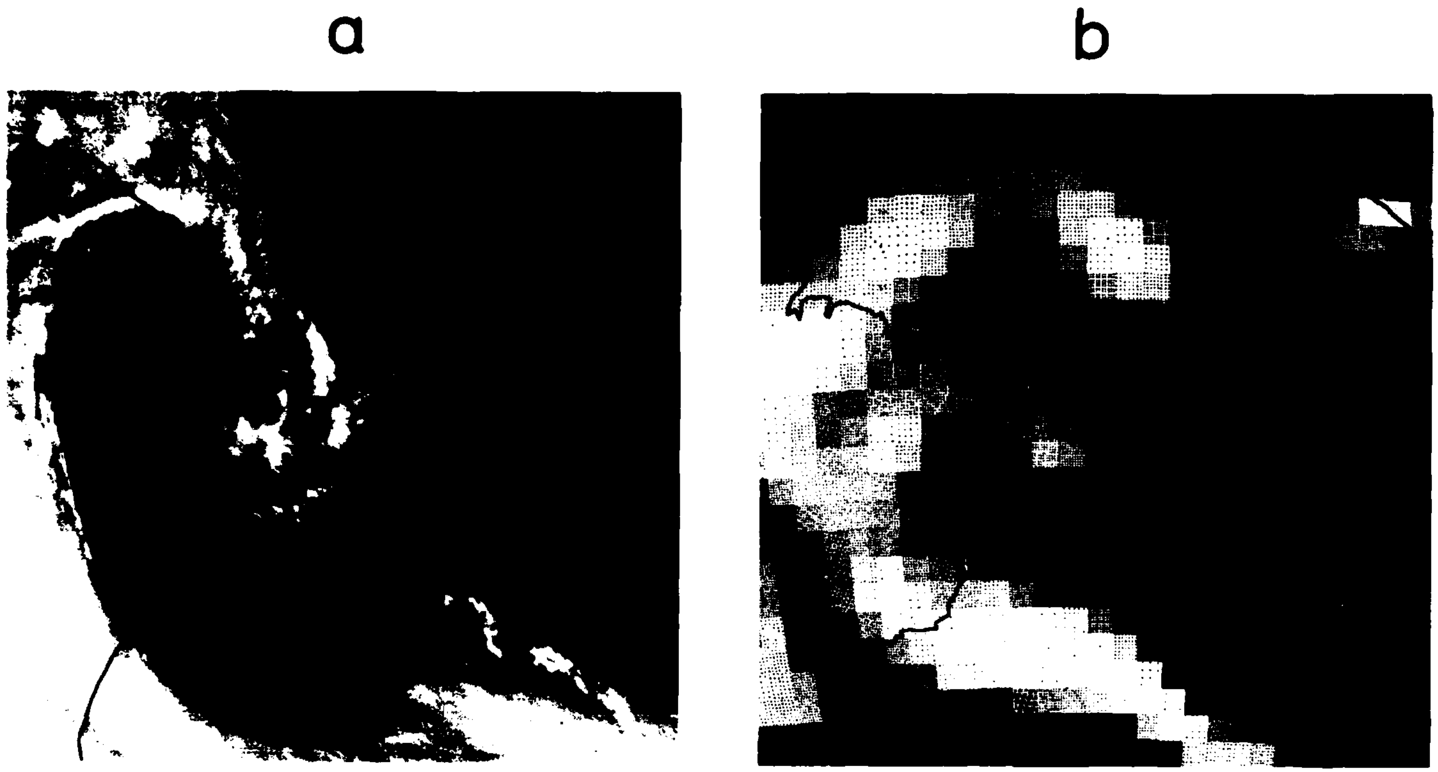

Fig. 3.2 Observed and model generated humidity fields. The METEOSAT water vapor channel image (a) and ECMWF model generated cloud cover (b) are shown over eastern Brazil and adjoining Atlantic at 1200 GMT 22 February 1979 (from Bengtsson et al., 1982b).

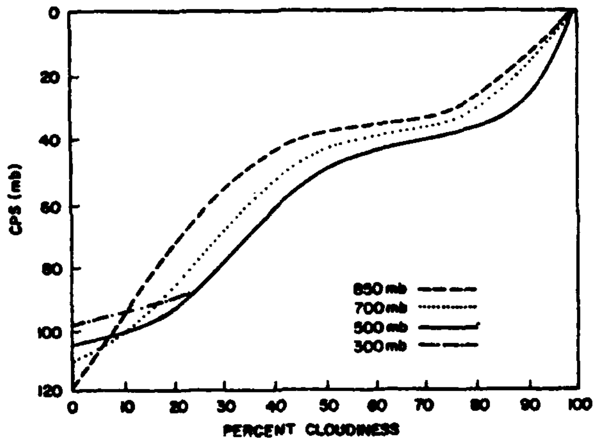

Fig. 3.3 Cloudiness versus humidity. The AFGWC empirical CPS-cloud conversion curves. The 850 mb and 300 mb curves serve also for 1000 mb and 400 mb respectively (from Mitchell and Warburton, 1983).

fields, especially their diurnal and latitudinal variations, may be quite different from those of plane-parallel continuous cloud. In light of our inadequate understanding of the physical mechanisms that control clouds, it may not be possible to parameterize finite cloud development and its effects. However, with the temperature information available from satellite data, some improvements (cf. WMO, 1978) in the treatment of cloud vertical extent become possible. These improvements include distinguishing air temperature at physical cloud boundaries from the effective radiation temperatures of thick clouds and allowing fractional cloudiness within a grid box, possibly with different temperatures in clear and cloudy parts of the column.

3.1.4 Cloud Phase

Cloud phase is not presently a model predicted parameter, except that some models distinguish cirrus (i.e., ice) cloud from water cloud for radiation calculation purposes (cf. Somerville et al, 1974; Hansen et al., 1983). The cirrus cloud is identified through an arbitrarily assigned "threshold temperature," i.e., if the cloud layer temperatures is less than the threshold temperature, then the cloud is identified to be cirrus.

Because information about the phase of water in clouds, especially high level clouds, may be inferred from satellite data, and because the radiative effect of ice clouds is very different from water clouds, it will be important for the model to carry this parameter explicitly. However, to do this, a complete treatment of the water budget, including vapor, liquid and solid phases, will be needed.

3.2 Radiation Parameterization

In this section, we briefly review the treatment of the model radiation field. A more detailed review can be found in the papers by Rodgers (see WMO, 1975) and by Cox (see WMO, 1978).

Generally speaking, clear atmosphere radiation parameterizations currently adopted in GCMs used for NWP have changed from the concept of precalculated solar reflectivity, absorptivity, and transmissivity, as well as thermal emissivity to more sophisticated radiative transfer schemes. However, representation of the cloud radiation field remains crude due primarily to the requirement of computational efficiency and to the inadequate knowledge of cloud radiative properties. As observational data become abundant, more sophisticated cloud radiation treatments are expected.

3.2.1 Input Parameters

In NWP models, the necessary input parameters in the radiation parameterization include (1) temperature and cloud distributions, (2) distributions of water vapor, ozone and carbon dioxide concentrations and (3) surface temperature, pressure and spectral properties.

Carbon dioxide concentration is usually fixed to be a value around 300-330 ppmv, and seasonal and latitudinal climatological values are used for ozone. Surface spectral properties are also specified. Therefore, water vapor and cloud, as well as temperature, are the model generated parameters that interact with the radiation field.

3.2.2 Clear Atmosphere

In NWP models, two types of radiation calculation have to be performed: solar radiation (the source of energy which drives the atmosphere) and thermal radiation (the radiation energy sink). Determination of the distribution of the radiation energy source and sink requires the solution of the radiative transfer equation which couples the processes of selective absorption and emission due to water vapor, carbon dioxide and ozone with multiple scattering due to air molecules (Rayleigh scattering) and particles (aerosols and cloud particles). The degree of the importance of these processes depends highly on the spectral regimes under consideration. The same processes are also involved in the retrieval problem. We discuss the physical basis of radiative transfer in the atmosphere in more detail in Section 4.3.3.

Solar radiation is absorbed and scattered by the atmosphere, the oceans and the land surfaces. In NWP model parameterizations, major processes that need to be included are: (1) absorption by water vapor, carbon dioxide and ozone, (2) scattering due to air molecules and (3) scattering and absorption by aerosol and cloud particles. Surface reflectivity is an important parameter because of the large differences between the ocean and land surfaces, and between the various types of land surfaces. Some models (e.g., Hansen et al., 1983) employ wind speed dependent ocean reflectivity.

Thermal radiation is emitted and reabsorbed by the atmosphere, clouds and surface. The important processes are (1) absorption and emission by water vapor, ozone, and carbon dioxide throughout the troposphere and stratosphere, and (2) absorption and emission by clouds. The surface emissivity in the 10 μm window is now often assumed to be less than unity (cf. Ramanathan et al., 1983; Hansen et al., 1983).

Because of the demand for computational efficiency in GCMs, simplifications are required in calculating the radiation field, particularly in: (1) the integration over frequency and (2) the treatment of the temperature and pressure effect of the inhomogeneous atmosphere on absorption and emission. Table 3.1, taken from WMO (1975), summarizes the commonly used methods to treat these problems. The most popular approach used in the early complex GCMs (e.g., NCAR, GFDL, GLAS) was the emissivity approximation (Manabe and Strickler, 1964; Sasamori, 1968; Lacis and Hansen, 1974). However, more recent radiation parameterizations have employed more sophisticated multi-interval methods for frequency integration (Fels and Kaplan, 1974; Hansen et al., 1983; Dickinson et al., 1978; Ramanathan et al., 1983).

3.2.3 Cloudy Atmosphere

Thermal Radiation

Low and middle level clouds are generally assumed in GCMs to be black, i.e. opaque to infrared radiation and their emissivity taken to be unity. This assumption may introduce substantial errors in some cases, e.g. in high latitudes during winter when the clouds are not opaque and are persistent (cf. Herman and Goody, 1976).

The treatment of high level cloud is also crude and can usually be categorized as one of two methods. The first is to assume the cloud to be either black or gray. For a gray high level cloud, a value of 50% is usually used for both the emissivity and transmissivity within the 8-12 μm window and the cloud is assumed to be black elsewhere. In this method the models do not distinguish ice cloud from water cloud. The second method is to specify ice crystal clouds if the layer temperature is less than -40 K (Somerville et al., 1974; Hansen et al., 1983), and then to use pre-assigned ice particle radiative properties for radiation flux calculations. For example, in the GISS GCM, the radiative properties of absorption and scattering coefficients are calculated based on the Mie theory using observed cloud particle size distribution, while the optical depth is determined from the visible values (see

Table 3.1 Methods of computing radiative transfer. The methods are in order of increasing accuracy and decreasing computational speed (from WMO, 1975).

Solar Radiation		Terrestrial Radiation	Notes
Scattering	Absorption		
Climatology	Climatology	Climatology	Non-interactive
		Newtonian cooling	Interactive with temperature
Empirical relations based on observations or calculations	Empirical relationships based on observations	Various empirical relation-	Interactive with temperature
	Integrated absorption for each gas separately	Emissivity, as in radiation charts. Scaling approximation	May be fully interactive
Mean photon path approximation	As terrestrial radiation	Many frequency intervals Curtis Godson approximation Band models	
Frequency integration of doubling method, etc.	Line by line integration	Line by line integration	Very slow

Solar Radiation below). Water clouds are considered to be black. These two treatments are still used in most GCMs, although work is underway to improve them. For example, Ramanathan et al. (1983) have recently incorporated a more sophisticated cloud radiation scheme into the NCAR GCM by allowing for interactions between thermal radiation and model predicted cloud condensed water content.

It should be noted that the functional dependence of radiative flux on emissivity and cloud fraction is only through their product (i.e., through effective cloud fraction $F_{eff} = F\varepsilon_c(\nu)$; see Eq. (4-10b) below), and that this is in fact the quantity obtained by physical retrieval from satellite data (Chahine, 1982).

Solar Radiation

Studies of stratiform cloud solar radiative properties and their parameterizations for use in GCMs are numerous because the stratiform clouds dominate the mid and high latitude radiation budget and because more observational data are available. The simplest and most common technique is to specify the reflectivity, absorptivity and transmissivity of clouds to solar radiation. These radiative properties are also pre-calculated using observed cloud particle size distributions. The cloud visible optical depth is determined to yield realistic cloud albedos. Therefore, the visible cloud optical depth is a model tuning parameter and its value depends on cloud type, altitude and thickness. Simple radiative transfer schemes are usually employed for radiative flux calculations (e.g., Lacis and Hansen, 1974; Schmetz and Raschke, 1979).

Most NWP models do not treat fractional cloudiness for radiation calculations, although recently this has been incorporated into some models (Schmetz and Raschke, 1979; Slingo, 1980). Recent improvement of solar radiation parameterization, like the thermal radiation parameterization for cirrus, is concentrated on the incorporation of interactions between the model predicted condensed water content and visible optical depth, and thus the radiation fluxes.

3.3 Boundary Layer Physics

The interaction between the atmosphere and the underlying surface results in a boundary layer where turbulence is the principal mechanism for air-surface momentum and moisture exchanges and, along with radiation, heat exchange. The small-scale turbulence is believed to be intermittent above the boundary layer, while the transport of moisture and heat to the upper atmospheric layers (from the boundary layer) is dominated by large-scale vertical motions and by deep cumulus convective processes. Therefore, in GCMs, parameterizations for the small-scale momentum, moisture and sensible heat exchanges are needed and these changes are to be evaluated on the basis of averaged values of wind, temperature and humidity generated within the models, and from values of surface properties such as surface roughness, soil moisture and temperature.

Approaches to boundary layer parameterization have traditionally been based on the use of similarity and dimensional considerations, such as the well-known Monin-Obukhov similarity for the surface layer of the atmosphere (cf. Monin, 1970). Deardorff (1972) has proposed a parameterization for the boundary layer, stressing that the height of the boundary layer is generally time-dependent and therefore its evolution must be considered. Recent progress in the parameterizations of the boundary layer in NWP models has been reviewed by Mahrt (1983). Here, we briefly describe these parameterizations from the point of view of improving them by using satellite derived information.

3.3.1 Fluxes of Sensible Heat and Moisture

The sensible heat (F_h) and moisture (F_q) fluxes are usually calculated in the GCMs based on the bulk aerodynamic formula (see Deardorff, 1967)

$$F_h = C_p \rho C_h V_s (T_g - T_s), \tag{3-3}$$

$$F_q = D_w \rho C_q V_s (q_g - q_s), \tag{3-4}$$

where the subscripts s and g stand for surface air and ground. The drag coefficients C_h and C_q for heat and moisture transfers are usually assumed to be the same. q_g is the saturation specific humidity of the ground temperature T_g, V_s the surface wind, D_w a wetness factor which sets the evaporation from the surface to be a specified fraction of the evaporation from a saturated surface, ρ the air density and C_p the specific heat of air at constant pressure. These parameterizations depend on the model calculated V_s, T_s, T_g, and

q_s (see below) as well as the empirically determined D_w and the drag coefficients.

The expressions (3-3) and (3-4) work very well over ocean. However, over land they are less satisfactory due to the inhomogeneous nature of the land surface (cf. Garratt, 1977).

3.3.2 Surface Air Temperature and Specific Humidity

The surface air temperature (T_s) and specific humidity (q_s) are calculated in the models by equating the fluxes of sensible heat [Eq. (3-3)] and water vapor [Eq. (3-4)] from the lower boundary to the fluxes into the lowest model atmospheric layers. For the latter,

$$F_h = C_p \rho K \left(\frac{T_l - T_s}{z_l - z_s}\right), \tag{3-5}$$

$$F_q = \rho K \left(\frac{q_l - q_s}{z_l - z_s}\right), \tag{3-6}$$

where T_l is the potential temperature of the layer, K an empirical diffusion coefficient (Deardorff, 1967) and z_s and z_l, the height of the surface layer and lowest model atmospheric layer respectively.

3.3.3 Surface Wind

The surface wind (V_s) is usually estimated by extrapolating the model calculated wind velocities in the two lowest atmospheric layers to the surface.

3.4 Ground Physics

3.4.1 Ground Temperature

Ground temperature at land and ice locations is computed using an energy balance equation that takes into account the solar and thermal radiation fluxes, the sensible and latent heat fluxes, the heat conduction, the melting of snow and ice, if present, and the ground heat capacity. A minimum of two ground layers is usually used, an upper thin layer to simulate the diurnal variations and a lower layer for seasonal heat storage.

In NWP models, ground temperature at ocean locations is usually specified from climatological sea surface temperatures.

3.4.2 Ground Moisture

The water content over the land surface is calculated in the models through a water balance equation. The rate of change of moisture (W) in the upper land surface layer is

$$\frac{\partial W}{\partial t} = P_r - E - R \tag{3-7}$$

where P_r, E and R are the rainfall, evaporation and total run-off rates.

The parameterizations of evaporation and run-off, which depend on the ground wetness, soil type and vegetation, are usually determined empirically from field measurements (cf. Sellers, 1965).

3.4.3 Snow and Ice

The equation of balance of snow is

$$\frac{\partial S}{\partial t} = P_s - E_s - S_m \tag{3-8}$$

where S is the snow cover amount in equivalent liquid water amount, P_s the snow precipitation rate, E_s the rate of sublimation, and S_m the snow melt rate. The decision as to whether rain or snow is falling is based on whether the temperature of the lower model layer above the ground is warmer or cooler than 0 C. If the ground temperature is less than 0 C, the snow depth increases; otherwise the snow melts, decreasing the surface temperature. Snow melt goes into runoff or ground moisture the same as the precipitation.

Sea ice is a product of dynamical and thermodynamical processes both in the ocean and atmosphere. The present treatment of ice melting and formation in the models is very simple. The change of ice thickness due to melting is calculated from the heat balance between precipitation, sublimation of ice surface and the turbulent heat flux between the ice surface and the atmosphere and ocean. Ice transport by the oceans is ignored.

3.5 Tuning and Verification

Tuning and verification have always been an integral part of developing model physical parameterizations. The usual procedure is to develop the parameterization either theoretically (e.g., the radiation parameterization) or empirically, based on observational data (e.g., simple cloud parameterizations using criteria such as (3-1)) and then to compare the model simulations (e.g., radiation flux, planetary albedo, cloud cover, etc.) with observations, changing (or tuning) the key adjustable parameters in the parameterizations so that better agreement can be obtained. Recent examples can be found in Slingo (1980), Ramanathan et al. (1983), and Hansen et al. (1983).

Gordon and others at GFDL have begun experiments in tuning their cloud parameterizations by constraining the low and high cloud amounts to be radiatively consistent with the observed net flux while specifying other cloud parameters, such as cloud tops, cloud emissivities and cloud albedos a priori (WGNE, 1983). These parameters could, in principle, also be predicted by the model and tuned against measurements by satellites. This does not seem to have been attempted yet, nor has use of retrievals of the moisture field.

3.6 Sensitivity and Impact Studies

The evidence for the importance of physical parameterizations comes mostly from sensitivity and impact studies. In most operational models the improvements in physical parameterizations have been evolutionary and it is usually not possible to identify specific significant impacts. However, recent experience at ECMWF, where the model physical parameterizations have been updated over relatively short intervals, indicates that improvements in the physics lead to better forecasts even for short range (L. Bengtsson, private communication). Their successes include good forecasts, from one to five days, of stratiform cloud patterns (Sundqvist, 1981).

Recently, impact studies of the radiation and cloud parameterizations have been conducted by Raschke and others at the University of Bonn (Geleyn and Morcrette, private communication). The results indicate that the atmospheric energetics are very sensitive to the model radiation parameterization even on the time scale of a few days and that consideration of cloud liquid water is essential for the correct treatment of the cloud-radiation feedback.

Some long-term forecast experiments were conducted at GFDL using different parameterizations of convection and the boundary layer (Miyakoda and Sirutis, 1977; Sirutis et al., 1980). The time and zonally averaged humidity fields in these experiments were very similar despite the difference in the parameterizations and major differences in the terms of the budget maintaining these fields.

A number of impact studies have addressed the importance of latent heat release. The importance of latent heating as an energy source for synoptic scale disturbances has been discussed by several authors (Aubert, 1957; Danard, 1964; Tracton, 1973; Kreitzberg, 1976; Anthes and Keyser, 1979; Perkey, 1980; Anthes, 1983). In particular, it is anticipated that intense explosively deeping extratropical cyclones depend on latent heat release as an energy source. These storms have been the focus of several diagnostic and forecasting studies for this and other reasons: They are difficult to forecast. The physical processes that drive them and their organization are scientifically interesting. They pose a threat to maritime interests. Sanders and Gyakum (1980) have defined a bomb as an extratropical storm which deepens at a rate of at least 1 mb h^{-1} for an extended period of time (e.g., 12 h). Sanders and Gyakum made an extensive study of observed bombs, identifying 267 northern hemisphere occurrences in a 3-year sample. They found some geographic correlation with strong sea surface temperature gradients. In the northern hemisphere, bombs are most frequent in winter in the western Pacific and Atlantic. Sanders and Gyakum concluded that realistic parameterizations of the boundary layer and cumulus convection are required to predict these storms. They found that the NMC 6-layer (6-L) primitive equation (PE) model with 381 km grid resolution systematically underpredicted the observed deepening of these storms by O(25 mb/day) while the 7-layer (7-L) PE model with a 190.5 km grid resolution underpredicted these storms by O(20 mb/day). Current models typically underpredict the deepening of bombs by O(15 mb/day) (Murty et al., 1983). The trend of these numbers suggest that high resolution may be required to accurately predict bombs, but Sanders and Gyakum (1980) and Gyakum (1983b) performed diagnostic calculations with a continuous quasi-geostrophic model which show that high resolution dynamics alone can not predict the observed intensification. The experiments mentioned below in this section and in Section 5 show that cases of cyclogenesis, especially those of very rapid cyclogenesis, are sensitive to the initial specification of humidity and the moist physical processes. (Most of these studies employed regional models.) The forecasts are generally improved when the initial conditions and parameterizations are made more realistic, but these improvements are not large when compared to the forecast error. The bomb problem is not yet solved, but the evidence suggests that improved parameterizations are necessarily part of the solution.

In several recent experiments with regional scale models, a control experiment with complete physics is compared to a dry model experiment (e.g., the study by Maddox et al. (1982) discussed in Section 1). These experiments have shown the importance of the parameterized latent heat release in predicting rapid cyclogenesis (Anthes et al., 1982, 1983; Chang et al., 1982) in the evolution of the environment just prior to cyclogenesis (Uccellini et al., 1983) and in the generation of eddy kinetic energy (Chang et al., 1983; Robertson and Smith, 1983; Kennedy and Smith, 1983). Similar experiments have also implicated other factors: Uccellini et al. (1983) also stress the sensitivity of these storms to sensible heating while Anthes et al. (1983) found sensitivity to the initial conditions, horizontal resolution and vertical resolution in the PBL. Juvanon du Vachat et al. (1983) show that the prediction of cyclogenesis is sensitive to the cumulus parameterization used.

The vertical distribution of latent heat release has a significant impact on cyclogenesis. When latent heat release is maximized at low levels, it tends to destabilize the atmosphere, and it must be accompanied by substantial convergence in the boundary layer which, at least, over the ocean, provides positive feedback (Gyakum, 1983b). Anthes et al. (1983) found substantial differences in predicted intensity were related to the vertical distribution of convective latent heating. This sensitivity is clearly shown in an experiment reported by Anthes and Keyser (1979). In these experiments, Anthes and Keyser prespecified the vertical distribution of convective latent heating. For the case of 0000 GMT 3 March 1978, their model overpredicted the development of a cyclone over the Florida panhandle by 16 mb. When the latent heating profile was shifted upwards the model no longer overpredicted the cyclogenesis. The vertical distribution of latent heating is also very important in the tropics for the development of tropical cyclones (Anthes, 1982; and references therein) and for its effect on the Walker circulation and on tropical-extratropical interactions (Hartmann et al., 1983).

Sensitivity studies of the model generated climate, to assess the importance of various physical parameters and their parameterizations, are very useful. Examples are studies of cloud-radiation interactions by Roach and Slingo (1979), Stephens and Webster (1979), Wetherald and Manabe (1980), Hunt et al. (1980) and Hunt (1982), and of cumulus parameterization by Stephens and Wilson (1980), Donner et al. (1982) and Bennetts and Rawlins (1981).

3.7 Concluding Remarks

The complexity of cloud interaction with atmospheric dynamics and radiation makes it very difficult to understand the cause-and-effect among these physical processes. However, the many observational and model studies briefly reviewed in this section have shed some light on this important problem. It appears that with the additional satellite information, improvements of the physical parameterizations used in large-scale numerical models can be expected. These are summarized below:

(1) A prognostic equation for cloud condensed water content is needed in order to link more realistically the model radiation field with clouds as well as to predict latent heat release and precipitation. Satellite-derived cloud liquid water information may then be used both as initial condition and for tuning and verification of cloud parameterization.

(2) It is quite clear from radiation calculations that the cloud phase is important. A distinction between water and ice clouds is necessary in the models.

(3) Cloud horizontal fraction coverage and vertical extent within a model grid box are both important parameters, and should be incorporated into the cloud parameterization scheme in such a way that those cloud characteristics, such as cloud topography, emissivity and albedo, which are measurable from satellites, are realistically sensitive to them.

(4) Satellite measurement of the geographical and temporal radiance distributions permits the verification of the model-employed clear sky radiation parameterization as well as the predicted surface characteristics (such as snow cover) and may be used for tuning and further improvement of cloud parameterization (cf. Gordon in WGNE, 1983). It is particularly important that studies along this line be actively pursued and encouraged.

(5) The atmospheric moisture comes from the underlying ocean and land surfaces. Consequently, the boundary layer, which serves as a buffer in regulating the upward movement of moisture, plays a critically important role. Further studies on the improvements of the model boundary layer parameterizations are needed, especially over the land surfaces.

4. REMOTE SENSING OF NWP RELATED GEOPHYSICAL PARAMETERS

Satellites are an integral part of our current atmospheric observing system. Without satellite data our view of the atmosphere would be severely limited. Most other data sources are land-based; commercial ships and aircraft provide some data over oceans but for the most part they follow the few main northern hemisphere transoceanic routes. The atmosphere of the southern hemisphere ocean regions would be virtually unknown without satellite data. The spacing of the conventional (i.e., non-satellite) data is rather coarse, typically 500 km. Conventional data are primarily available only at the synoptic times, and it is only at the main synoptic times, 0000 and 1200 GMT, that this coverage is complete. Satellite data are vital for filling the voids, both in space and in time, between the conventional data.

In addition to differences in coverage, satellite and conventional data differ in other fundamental ways, for example, in which quantity is measured and how it is sampled. These differences are discussed in detail by Pierson (1983) for the case of surface winds. Conventional measurements tend to be rather direct whereas meteorological satellite instruments measure intensity of electromagnetic radiation. Retrieval of geophysical parameters from satellite measurements is an inverse problem. In principle, the measured radiances may be directly assimilated since methods like optimal interpolation may make use of any observation which is correlated to the variable being analyzed. This possibility is discussed further below (Section 5.2) but current numerical weather prediction (NWP) practice is to separate the retrieval and assimilation problems.

Conventional data sources generally provide values that are nearly instantaneous or short (O(1 min)) time averages. Satellite data are nearly instantaneous averages over horizontal areas which range in size from O(1 km) to O(100 km). Satellite sensors also average in the vertical direction, but in a rather complicated way, as we will see in considerable detail below.

4.1 Overview

In this section we review the various measurements currently or soon to be available from satellites. These data include temperature profiles, single-level winds at the ocean surface and at cloud tops, water vapor profiles, vertically integrated water vapor amounts, a host of cloud parameters and surface properties such as albedo, surface temperature and soil moisture. Of these data, it is only the temperature profile data which have found widespread use in NWP. Because model resolution is so coarse relative to the density of satellite observations, current global NWP applications use only a fraction of the available temperature profiles. In the future, as model resolution becomes finer we expect that the full potential of satellite data will be exploited. The high horizontal resolution of satellite data will be especially important for describing the moisture and cloud fields, since moisture (and even more so, cloudiness) varies more and on smaller scales than other meteorological variables (Section 5.2).

Proposals for the remote inference of atmospheric structure from orbiting satellites (King, 1956; Kaplan, 1959, 1961) were motivated from a meteorological perspective. Early studies investigating the utility of incorporating remotely sensed temperature data in numerical models of the global circulation suggested significant potential (Charney et al., 1969; Halem and Jastrow, 1970). It is interesting to note, however, that in the decade and a half since the first successful retrieval of an atmospheric temperature profile using the NIMBUS 3 SIRS instrument (Wark and Hilleary, 1969), a consensus has only recently emerged regarding the usefulness of satellite data in NWP (Section 5.6). To a certain extent, it has been recognized that the apparent ambiguity of results reported by different groups is due to the "system dependency" of satellite sounding impact. For the most part, the system has been defined to consist of analysis and forecast components. Recent studies, however, have introduced questions concerning the limitations of the retrieval methods themselves and their effect on satellite sounding impact tests (Phillips et al., 1979; Schlatter, 1981; Phillips, 1981; Gruber and Watkins, 1982; McMillin and Dean, 1982; Koehler et al., 1983). For example, observed error statistics for satellite data currently used by ECMWF have been compiled by Bengtsson et al. (1982) and are reproduced in Table 4.1. These errors are for retrievals which use regression techniques. Better retrievals can be obtained by the use of physical methods, using regression retrievals as first guesses (J. Susskind, private communication). These considerations imply that retrieval methods are important NWP system components. We expect that the retrieval process will become part of the assimilation cycle in the future, with the forecast model providing a priori information for the retrievals.

Table 4.1 Observation errors for different observing systems. Sea surface pressure: SYNOP/SHIP 1.0 mb; buoy 2.0 mb. COLBA/DROPWINDSONDE/TWOS-NAVAID observation error is calculated from the level IIB quality information. Temperatures given as layer means (from Bengtsson et al., 1982a).

	Temperature			Wind ms^{-1}			
		TIROS-N		Radiosonde	Cloud Drift Wind		
Level (mb)	Radiosonde	Clear/ Partly Cloudy	Micro-Wave	Radiosonde Pilot ASDAR AIDS	NESS WIS-CONSIN	ESA LMD AIREP	HIMA-WARI
10	4.5	2.8	2.8	6	8	8	13
20	3.8	2.6	2.7	6	8	8	13
30	3.2	2.5	2.6	6	8	8	13
50	2.7	2.4	2.5	6	8	8	13
70	2.3	2.2	1.4	6	8	8	13
100	2.1	2.0	1.6	6	8	8	13
150	2.1	2.0	1.7	6	8	8	13
200	2.0	1.9	1.8	6	8	8	13
250	1.8	1.9	1.9	6	8	8	13
300	1.6	1.8	2.0	6	8	8	13
400	1.5	1.8	2.2	5	7	8	10
500	1.2	1.7	2.2	4	7	8	10
700	1.1	1.8	2.5	3	5	8	6
850	1.1	2.0	3.9	2	4	7	6
1000	-	-	-	2	4	7	6

Work along these lines is currently underway at GLAS (J. Susskind, private communication).

In this section we will briefly review inversion methods used in atmospheric remote sensing. Comprehensive reviews of remote sensing of geophysical parameters have been provided by Derr (1972), Lintz and Simonett (1976), Lhermitte (1979), Deepak (1977, 1980), Chahine (1980), Harries (1980), and Gage and Balsley (1983). Current satellite data availability is summarized in Section 4.2. The fundamental forward and inverse problems of remote sensing will be discussed in Section 4.3. We will then discuss retrieval of global temperature (Section 4.4) and winds (Section 4.5). In order to support the goal of improving short-range prediction of hydrological variables, such as cloud cover, the primary focus of the discussion below will be on analyses of moisture related variables (i.e. humidity, cloud liquid water, precipitation) and bulk cloud parameters (i.e. cloud coverage fraction, cloud top height, etc. (Section 4.6)). Finally, surface variables that are either active in the hydrological cycle or contributing to atmospheric heating through the surface fluxes of latent and sensible heat or albedo effects will be covered in Section 4.7. These variables include snow and ice cover, soil moisture and surface temperature.

4.2 Current Satellite Data Availability

Meteorological data are available from satellite systems of NOAA NESS. These data are archived by and may be obtained from the Satellite Data Services Division (SDSD) of NOAA NCC. The available digital data (as well as images and movies) are summarized by Dismachek et al. (1980). Included in this summary are data collected by instruments on both geostationary and polar orbiting types of satellites. The geostationary satellites of the SMS/GOES series are in geosynchronous orbit at 35.8 x 10^3 km over the equator and carry a Visible and Infrared Spin Scan Radiometer (VISSR), (Ludwig, 1975), whose visible and infrared horizontal data resolution are approximately 1 km and 8 km respectively. Polar orbiting satellites of the TIROS-N series are equipped with the five-channel Advanced Very High Resolution Radiometer (AVHRR) imaging device and the TIROS Operational Vertical Sounder (TOVS) containing the High Resolution Infrared Sounder (HIRS/2), the Stratospheric Sounder Unit (SSU) and the Microwave Sounder Unit (MSU). (See Section 4.4.1 for a detailed description of these instruments.)

Low, middle, and high level winds are determined by analysis of cloud motion vectors which are observed by the geostationary satellites. Low-level wind vectors are computer-derived from photographic images with a sampling resolution of about 250 km. Middle and high-level winds are produced by a man-machine interactive system (Bristor, 1975). The winds from cloud motions at the analysis times of 0, 12 and 18 GMT are available on teletype at 3, 15, and 21 GMT. The wind accuracy is comparable to that of balloon observations (Hubert, 1976).

Atmospheric soundings are provided from the instruments of the TOVS system. Some 16,800 individual soundings are produced per day by the two NESS satellites and these are processed into values of mean temperature and precipitable water for layers between standard pressure levels (Modali and Novak, 1979). These, as well as the radiances of the MSU, SSU, and HIRS/2 are archived by the SDSD. Some of these sounding products are transmitted over the Global Telecommunications Service (GTS) network.

Parameters of the global heat budget are measured by the AVHRR. Long wave flux (on a 24 h basis) and absorbed solar radiation (for day time only) are measured with an accuracy of 7 Wm^{-2} with a spatial resolution of approximately 100 km (Dismachek et al., 1980). These data are archived on a 2.5° latitude-longitude grid. In addition, global sea surface temperature (SST) observations are derived from the AVHRR sensor by the SDSD. Selected SST observations are available through the GTS. Analyses of the SST measurements are created by NOAA on a daily basis for large scale 5° and 1° grids and, on a weekly basis, for regional 0.5° grids. Basic SST observations have an accuracy goal of ±1.5°C. This goal has been reached over approximately 70% of the ocean (Dismachek et al., 1980).

Data from the DMSP are not usually archived.

4.3 Retrieval Methods

With the exception of broad band radiative flux measurements which can be used in radiative budget calculations, the radiances measured by satellites is not of direct geophysical interest. The measurements may be expressed mathematically as complicated convolution integrals involving the parameters of interest. The inversion of these measurements is known as the retrieval or inverse problem.

4.3.1 The Inverse Problem

The fundamental retrieval problem is succinctly stated: given a set of measurements of the radiation emitted, scattered, and/or reflected from the earth-atmosphere system what can be deduced about the physical state of this system? The response to this question is an amalgam of mathematical elegance, computational brute force, and operational expediency. There are many individual retrieval methods. However, a recurrent theme of the many comprehensive reviews available (Westwater and Strand, 1972; Fritz et al., 1972; Smith, 1972; Fymat, 1975; Rodgers, 1976; Twomey, 1977; Parker, 1977; Deepak, 1977; Fymat and Zuev, 1978) is the conceptual unity among apparently diverse approaches. This is due to the fact that the underlying mathematical problem is always the same.

Stated symbolically, we are given measurements of radiance d_j, $j = 1,\ldots M$ which are the sum of an unknown error, ε_j and a contribution F_j which depends on the geophysical parameters u_i, $i = 1,\ldots N$. That is

$$d_j = F_j(u_1, u_2 \ldots u_N) + \varepsilon_j \qquad j = 1,\ldots M. \tag{4-1}$$

The F_j are known as mapping functions. The retrieval problem is: given the d_j determine the u_i. For the problem at hand the F_j are determined by the wavelength dependent radiative transfer processes, including absorption, emission and scattering by the atmosphere, by clouds and by the surface. The geophysical variables u_i include atmospheric temperature T, surface pressure p_s, and water vapor amounts. For the purpose of discussion consider a mapping function of the form

$$F = \int_0^{\ln ps} B(T(p))\, W(p,T(p))\, d\ln p$$

where B, the Planck function, and W, the weighting function, are known. The major contribution to the mapping functions for the temperature sounding channels has this form. Since T(p) is a function of p for $0 < p < p_s$ but the number of channels, M, is finite, such problems, even in the absence of measurement errors, do not have unique solutions and are thus not well-posed. To be <u>well-posed</u> a problem must have a solution which exists, which is unique and which depends continuously on the data of the problem. By <u>data</u> we mean the parameters, functions, etc., used to specify the particular problem. In our example, the measurements d_j and the functions B and W are the data. The retrieval problem may be rendered well-posed by adding some constraints on the physical parameters to be retrieved. For example, T(p) may be expressed as a finite sum of known empirical structure functions. With this formulation the coefficients of the structure functions become the u_i and the problem reduces to a finite set of algebraic equations. Such systems of equations will not in general be well-posed, but if the number of (independent) channels is equal to the number of coefficients the problem will be well-posed. However, well-posedness and its implied continuous dependence on the data is not enough to insure that the solution to a problem will be useful when the data are uncertain. In practice, the data are always uncertain, and solutions of a well-posed problem can still be ill-conditioned, that is very sensitive to the data. When this is the case small measurement errors may be translated into large errors in the retrieved variables. Ill-conditioning occurs for example when variation of one of the geophysical parameters has only a small effect on the measured radiances. In our example, decreasing the number of coefficients used and minimizing the sum of squared ε_j will usually improve the conditioning of the problem.

Ultimately, all retrieval methods are constructed to address these considerations, ill-posedness and ill-conditioning. One or both of these attributes may cause a particular treatment to produce spurious (i.e., non-physical) results or fail to converge (instability). Methods which are successful circumvent these potential pitfalls by introducing some (explicit or implicit) constraint on the solution based on a priori knowledge of the physical system. We will classify the constraints as one of the following: (1) a smoothness constraint, (2) a statistical constraint, or (3) a physical constraint. These classes of constraints correspond to three types of retrieval methods: smoothing (Phillips, 1962; Twomey, 1963; Tikhonov, 1963) as applied, for example, by Wark and Fleming (1966); statistical (Rodgers, 1966, 1971; Strand and Westwater, 1968; Wahba, 1969; Franklin, 1970; Smith et al., 1970, 1972; Turchin et al., 1971; Gaut et al., 1972; Smith and Woolf, 1976); and physical (Chahine, 1968, 1970, 1972, 1974, 1977, 1983; Chahine et al., 1977; Smith, 1970; Barcilon, 1975; Twomey et al., 1977). A fourth approach (Backus and Gilbert, 1967, 1968, 1970; Backus, 1970; Westwater and Cohen, 1973; Newman, 1979) which is usually applied to the design and analysis of the instruments themselves rather than to the actual retrievals, recognizes that sufficient constraints may not exist to render the problem well-posed and, furthermore, that it may not be possible to reduce noise levels in the data to a level which results in acceptable conditioning of the retrieval problem. Thus, there is a tradeoff between acceptable noise and sufficient resolution in retrieving the desired parameters. The approach of Backus and Gilbert provides criteria to select the set of retrieved parameters. This approach has been applied diagnostically to study the vertical resolution available from temperature retrievals (Conrath, 1972; Thompson, 1982).

Although all the techniques cited above are designed to solve the same mathematical problem, experience has proven that they may differ considerably in practice. This is a consequence of using different a priori data, adaptability to a particular problem, and computational details. This point is addressed in a critical assessment of four inversion techniques, one representative of each of the above retrieval methods applied to a single problem by Wolfson et al. (1979a, 1979b).

4.3.2 The Forward Problem

Often overlooked in discussions of inverse methods is the fundamental importance of the associated direct or forward problem symbolized by the mapping function, F in Eq. 4-1. A thorough understanding of the forward problem is essential to the successful integration of satellite data into a numerical forecast model.

In general our ability to infer a change in a geophysical parameter of magnitude δu_i will depend on the detection of a measurable change in the signal δd_j of one of the satellite sensor channels. These increments are related through the mapping function F by:

$$\delta d_j = F(u_j + \delta u_j) - F(u_j) \qquad (4\text{-}2)$$

If the signal is not detectible, say above the sensor noise level n_j, i.e. if

$$\delta d_j / n_j = O(1) \qquad (4\text{-}3)$$

then the parameter u_j is indeterminate in practice and this contributes to the ill-posedness of the problem. Sensitivity analyses based on (4-2) allow one to optimally select the channels needed to infer the desired geophysical parameters. The forward problem is also used in simulation studies of retrieval methods.

The forward problem also provides a direct test of many of these basic physical mechanisms (e.g., gaseous absorption, scattering by clouds) which we desire to include in the parameterizations of the prediction model itself (Section 3.2). By comparing measured satellite radiance data with corresponding simulated data, deficiencies in our basic understanding of the forward problem can be identified. This source of feedback should be most important for physical retrieval methods.

Doubts have been raised concerning the capability of performing the forward problem radiative transfer calculation within our current state-of-the-art knowledge of atmospheric transmittances and modeling of surface emission properties. Studies by McClatchey (1976) and Valovcin (1981) found positive biases (i.e., simulated radiances larger than observed radiances) in comparisons based on the DMSP 15 μm carbon dioxide and 20 μm water vapor channels and collocated radiosondes. Similar biases were reported by Weinreb (1979) and by Liou et al. (1980a) for the HIRS and Liou et al. (1980a) for the DMSP/ SSH and SSM/T. This latter study did not indicate discrepancies for the NIMBUS 6 SCAMS. Forward problem discrepancies such as these have been cited as limitations on the use of physical retrieval methods. However, Susskind et al. (1983) have recently obtained retrievals from HIRS2 and MSU radiances in clear conditions in simulation with accuracies of 0.7 and 1.0° K, respectively, using a physical method with a finely tuned empirical transmission model.

4.3.3 Physical Basis

The dependence of remotely-sensed data on desired atmospheric parameters (i.e., the forward problem) by equation (4-1) can be stated rigorously in terms of an appropriate solution to the radiative transfer equation. Stated generally, the satellite-incident radiance in the ith sensor channel (with characteristic wavelength, ν_i for look angle θ), $R_i(\theta)$, is given by (Goody, 1964; Kondratyev, 1969):

$$R_i(\theta) = R_i(p_b)\ \tau_i(\theta,p_b) + \int_{p_b}^{0} S(p)\ \frac{d}{dp}\ \tau_i(\theta,p)\ dp \qquad (4\text{-}4)$$

where $R_i(p_b)$ is the radiance emerging from the level at pressure p_b, $\tau_i(\theta,p)$ is the mean atmospheric transmittance evaluated from level p to space along look angle θ, and S(p) is the source function of emitted and/or scattered radiance at level p in the atmosphere. Usually p_b is a physical boundary such as the Earth's surface or the top of a cloud.

The atmospheric transmittance $\tau_i(\theta,p)$ consists of contributions by lines, τ_{il}, the continuum, τ_{ic}, and particulate extinction, τ_{ie}:

$$\tau_i(\theta,p) = \tau_{il}(\theta,p)\ \tau_{ic}(\theta,p)\ \tau_{ie}(\theta,p)\ . \tag{4-5}$$

The contribution due to n discrete absorption lines is

$$\tau_{il}(\theta,p) = \int \phi_i(\nu)\ \exp \frac{\sec\theta}{g} \sum_n \int_p^0 k_n[\nu,p',T(p')]\ q_n(p')\ dp'd\nu \tag{4-6}$$

where ϕ_i is the response (i.e., filter) function of the ith sensor channel, $k_n(\nu,p,T)$ is the absorption coefficient for the nth line, and $q_n(p)$ is the mixing ratio profile of the absorbing gas contributing to the nth line. The absorption coefficient depends on an assumed spectral line shape (see Goody, 1964) and an appropriate set of absorption line parameters (Rothman and McClatchey, 1976; Rothman, 1978, 1981; Rothman et al., 1982; Poynter and Pickett, 1980). Much of the inaccuracy in calculating the forward problem may be traced to uncertainties in determining these parameters. Existing techniques to evaluate (4-6) are described by McMillan and Fleming (1976, 1979), Fleming and McMillan (1977), Susskind and Searl (1978), McMillan et al. (1979), Clough et al. (1981), and Susskind et al. (1983).

Continua contributions to total transmissivity, τ_{ic}, include the 4 µm collision-induced absorption by nitrogen (see Kneizys, et al., 1980) and water vapor continua throughout much of the spectrum (Clough et al., 1980; Burch and Gryvnak, 1980; Liebe, 1980).

Extinction by particulates, $(1-\tau_{ie})$, includes the effect of Mie scattering (van de Hulst, 1957; Deirmendjian, 1969; McCartney, 1976), of aerosols in the visible and infrared (Shettle and Fenn, 1979), and of cloud and rain in the infrared and millimeter/microwave regions (Deirmendjian, 1975; Falcone et al., 1979).

The specific forms of $R_i(p_b)$, $\tau_i(\theta,p)$, and S(p) in equation (4-4) above will vary from one spectral region to another due to the wavelength dependence of atmospheric scattering and absorption processes. Consequently, the choice of observational wavelengths provides a means to focus selectively on particular geophysical parameters. Table 4.2 summarizes the features of surface and atmospheric source functions, $R_i(p_b)$ and $S(p)$, respectively, in different wavelength regions.

Parameters determining the radiometric properties of the surface (temperature, emissivity, albedo) and those related to these properties (e.g., surface wind over the ocean and soil moisture over land in the microwave region) are observable through the first term in (4-4) if the atmospheric contributions from the second term are negligible. Spectral regions satisfying these criteria are commonly referred to as atmospheric transmission "windows" and are used for imaging sensors such as the TIROS-N AVHRR and DMSP OLS and SSM/I. Window spectral regions are summarized in Table 4.3 from Fraser and Curran (1976). In addition to surface related information, imaging wavelengths also provide a means to observe the influence of atmospheric scatterers such as aerosol, cloud, and precipitation (depending on spectral region) on satellite incident radiances. In this case contributions from the second term in the r.h.s. of (4-4) are no longer negligible. Parameters potentially observable in this manner include cloud optical depth at visible wavelengths and cloud liquid water content and rainfall rate in the millimeter/microwave regions. In Eq. (4-4) the scattering source function contribution to the total source function (see Table 4.2) is given by the product of the single scattering albedo $\omega_i(p)$, i.e., the local ratio of total scattering to total extinction, and the scattering source function $J_i(p,\Omega)$ given by (Chandrasekhar, 1960; Kourganoff, 1963; Sobolev, 1963, 1975):

$$J_i(p,\Omega) = \frac{1}{4\pi} \int P_i(p,\Omega,\Omega') \; R_i(p,\Omega') \; d\Omega' \qquad (4\text{-}7)$$

where P_i is the angular scattering function and $R_i(p,\Omega')$ is the value of the local radiance field. Angular scattering (e.g. phase) functions for aerosol in the visible and infrared (Ridgway et al., 1982); cloud in the infrared (Yamamoto et al., 1971); and precipitation in the microwave (Savage, 1978) have been evaluated. Since the scattering source function depends on the local radiance field a numerical solution is required (Hansen and Travis, 1974; Shettle and Kuriyan, 1975; Lenoble, 1977; Fouquart et al., 1980).

Table 4.2 Radiative source functions. Forms of surface, $R_i(p_b)$, and atmospheric, $S(p)$, contributions to satellite-incident radiances, $R_i(\theta)$.

Spectral Region	Surface, $R_i(p_b)$	Atmosphere, $S(p)$
(1) Ultraviolet/visible ($\lambda < 0.7$ µm)	$\pi F_i \; \rho_i(\Omega_o) \; \tau_i \; (\theta_o, p_b)$	$\omega_i(p) \; J_i(p,\Omega)$
(2) Near infrared ($0.7 < \lambda < 4.0$)	$\pi F_i \rho_i(\Omega) \; \tau_i(\theta_o, p_b)$ $+ \varepsilon_i \; B_i \; [T(p_b)]$ $+ (1-\varepsilon_i) \; R_i{\downarrow}(p_b)$	$[1-\omega_i(p)] \; B_i[T(p)]$ $+ \omega_i(p) \; J_i(p,\Omega)$
(3) Infrared ($4.0 < \lambda < 100$ µm)	$\varepsilon_i B_i \; [T(p_b)]$ $+ (1-\varepsilon_i) \; R_i{\downarrow} \; (p_b)$	$B_i \; [T(p)]$
(4) Millimeter/microwave ($\lambda > 1000$ µm)	$\varepsilon_i B_i \; [T(p_b)]$ $+ (1-\varepsilon_i) \; R_i{\downarrow} \; (p_b)$	$[1-\omega_i(p)] \; B[T(p)]$ $+ \omega_i(p) \; J_i(p,\Omega)$

Legend:

F_i	solar irradiance
ρ_i	bidirectional reflectance of surface
Ω_o, θ_o	sun/sensor reflection angle, solar zenith angle
ε_i	surface emissivity
ω_i	single scattering albedo
$R_i{\downarrow}$	downward atmospheric flux
B_i	Planck function (thermal source function)
$J(p,\Omega)$	scattering source function for scattering angle Ω

Table 4.3 Transmission windows. Major atmospheric windows available for spacecraft remote sensing (from Fraser and Curran, 1976).

Ultraviolet and Visible	0.30-0.75 μm
	0.77-0.91
Near-Infrared	1.0-1.12
	1.19-1.34
	1.55-1.75
	2.05-2.4
Mid-Infrared	3.5-4.16
	4.5-5.0
Thermal Infrared	8.0-9.2
	10.2-12.4
	17.0-22.0
Microwave	2.06-2.22 mm
	3.0-3.75
	7.5-11.5
	20.0+

In spectral regions where scattering can be ignored (i.e., $\omega_i \sim 0.0$), the wavelength dependence of transmissivity can be exploited as a means to sound the atmosphere's vertical structure (Kaplan, 1959). Assuming unit surface emissivity ($\varepsilon_i = 1.0$) and choosing spectral regions with a single active absorber, Eqs. (4-4) and (4-6) may be combined using Table 4.2 to yield the following retrieval equation for the infrared and microwave regions:

$$R_i(\theta) = B_i[T(p_b)] \; \tau_i(\theta,p_b) + \frac{1}{g}\int_{p_b}^{0} B_i[T(p)] \; q(p) \; k(\nu_i,p) \; \exp[\frac{1}{g}\int_{p}^{0} q(p') \; k(\nu_i,p')dp']dp \tag{4-8}$$

For uniformly mixed gases such as CO_2 and O_2, q(p) is constant, and (4-8) may be used to obtain the temperature profile, T(p). Conversely, given the temperature profile, a mixing ratio profile (such as that of H_2O, for example) may be obtained. These retrievals are performed by applying the principles discussed above. Applicable sounding wavelengths and corresponding species are summarized in Table 4.4.

4.4 Temperature

4.4.1 Measurement Concepts and Implementation

Temperature is remotely sensed by measuring atmospheric thermal emission in absorption bands of molecules with known vertical distribution. Sounding of the atmospheric temperature profile is accomplished by selecting a set of spectrally distributed observational wavelengths with varying atmospheric transmittance. The more transparent regions (i.e., $\tau_i \sim 1.0$) sense deeper into the atmosphere than do opaque regions (i.e., $\tau_i \sim 0.0$) [cf. Eqs. (4-5), (4-6)]. If the thermal emission originates from an atmospheric constituent of known (preferably uniform) vertical distribution, the radiance observed may be qualitatively associated with an effective temperature in a specified atmospheric layer through the thermal source (i.e., Planck) function. Two gases in the Earth's atmosphere, carbon dioxide (CO_2) and oxygen (O_2) have both uniform mixing ratios and suitable emission features in spectral regions which can be used for temperature determinations from satellite-borne sensor systems (see Table 4.4). The temperature sounding equation may be expressed as (cf. Eq. (4-8), Table 4.2, Section 4.3.3):

Table 4.4 Sounding wavelengths and species.

Band	Species	Wavelength
Visible	O_3	0.6 μm
Near IR	H_2O	0.94 1.16 1.38 1.89
	CO_2-H_2O	2.51
Middle IR	CO_2	4.30
	H_2O	6.70
Thermal IR	O_3	9.6
	CO_2	15.0
	H_2O	~20
Microwave	H_2O	0.16 cm (183 GH_z)
		1.35 cm (22.235 GH_z)
	O_2	0.25 cm (118 GH_z)
		0.5 cm (60 GH_z)

$$R_i(\theta) = B_i[T(p_b)]\ \tau_i(\theta,p_b) + \int_{p_b}^{0} B_i[T(p)]\ W_i(\theta,p)\ d\ln p \qquad (4\text{-}9)$$

where $W_i(\theta,p) = d\tau_i/d\ln p$ is the "weighting function" for the ith spectral interval. The weighting function indicates the contribution of a particular level's temperature (emission) to the weighted radiance measured at the sensor. For a gas of known constant vertical distribution [q(p) = const] and assuming little temperature dependence of absorption coefficient, the weighting function may be calculated a priori. If a weighting function were strongly peaked in the vicinity of a single pressure level (say approaching a δ function) a single measurement would unambiguously provide the temperature at that level. Unfortunately, physical weighting functions are rather broad.

The concept of using narrow channels in the 15 µm CO_2 vibration-rotational bands (Kaplan, 1959) was experimentally tested aboard the NIMBUS 3 spacecraft launched in April 1969 using the Satellite Infrared Spectrometer (SIRS) instrument (Wark and Hilleary, 1969). The first instrument designed to provide temperature profiles operationally was the NOAA 2 Vertical Temperature Profile Radiometer (VTPR) launched in 1972 (Jastrow and Halem, 1973).

In the microwave spectral region, Meeks (Meeks, 1961; also Meeks and Lilley, 1963) suggested use of the 5 mm (60 GHz) spin-rotational band of O_2. Further microwave remote sensing concepts are discussed by Staelin (Staelin, 1966, 1969, 1981, and Njoku, 1982). A primary advantage of microwave over infrared sensors is the relative transparency of clouds at longer wavelengths (see Toong and Staelin, 1970). Temperature retrievals in the 60 GHz region are thus relatively unaffected by most cloud types (Staelin et al., 1975a, Liou et al., 1981); however, surface emissivity effects over land are a problem and the weighting functions tend to be broader than in the infrared. The first experimental space-borne microwave sounder was the NIMBUS E Microwave Spectrometer (NEMS) aboard NIMBUS 5 (Staelin et al., 1973). Its application to temperature retrievals is discussed by Waters et al. (1975). Potential use of the single O_2 transition at 2.5 mm wavelength (118 GHz) has been suggested (Staelin et al., 1978; Ali et al., 1980). Weighting functions are sharper but the effects of non-precipitating clouds are not negligible in this spectral region.

Because of the strong temperature dependence of black body radiance near 4 µm, the 4.3 µm CO_2 band and 3.7 µm window are especially useful in obtaining retrievals of tropospheric and surface temperatures (McClatchey, 1967; Kaplan, 1969; Shaw et al., 1970).

Of the temperature retrieval concepts described above, applications of the 4.3 and 15 μm CO_2 bands and the 5 mm O_2 complex have been implemented in the current (now third) generation civilian operational sounding system for polar orbiting satellites called the TIROS-N Operational Vertical Sounder (TOVS) consisting of three instrument subsystems (see Smith et al., 1979, 1981a). (1) The HIRS-2 is the second version of the High Resolution Infrared Radiation Sounder. HIRS-2 has 20 channels, including 7 channels in the 15 μm CO_2 band, 5 channels in the 4.3 μm CO_2 band, 3 water vapor channels, an ozone channel and 4 window channels. One of the window channels is in the visible. (2) The MSU is the Microwave Sounding Unit. The MSU has one window and 3 temperature channels. (3) The SSU is the Stratospheric Sounding Unit. The SSU has 3 channels in the 15 μm CO_2 band.

Weighting functions for the TOVS sensor subsystems are illustrated in Fig. 4.1 (from Smith et al., 1979). Note that in the lower and middle troposphere the vertical sampling attained in the 4.3 μm and 15.0 μm infrared CO_2 temperature channels is considerably sharper than that attained in the microwave O_2 channels. Conversely in the upper troposphere and stratosphere (say above 300 mb) the microwave weighting functions are narrower than those for the infrared channels. This suggests a complementary capability. The TOVS instrument system represents approximately 1976 state-of-the-art. TIROS-N was launched October 1978.

Analogous infrared sounding capabilities for temperature (and moisture, see Smith and Zhou, 1982) are implemented aboard the GOES geosynchronous satellites using the VAS (VISSR Atmospheric Sounder) instrument (Smith et al., 1981b). No microwave capability is currently available for geosynchronous satellites. Within the Defense Meteorological Satellite Program (DMSP) of polar orbiters, infrared temperature profilers employing the 15 μm CO_2 bands include the early special sensors designated SSE, its follow-on the SSH which first became operational in April 1977 (Savage, 1980) and the subsequent SSH/2. The DMSP microwave radiometer (SSM/T) is a seven frequency device (50-60 GHz, O_2) providing three stratospheric, three tropospheric, and one surface weighting functions (Rigone and Stogryn, 1977). Current plans call for an all microwave sounder system after the flight of the last SSH/2.

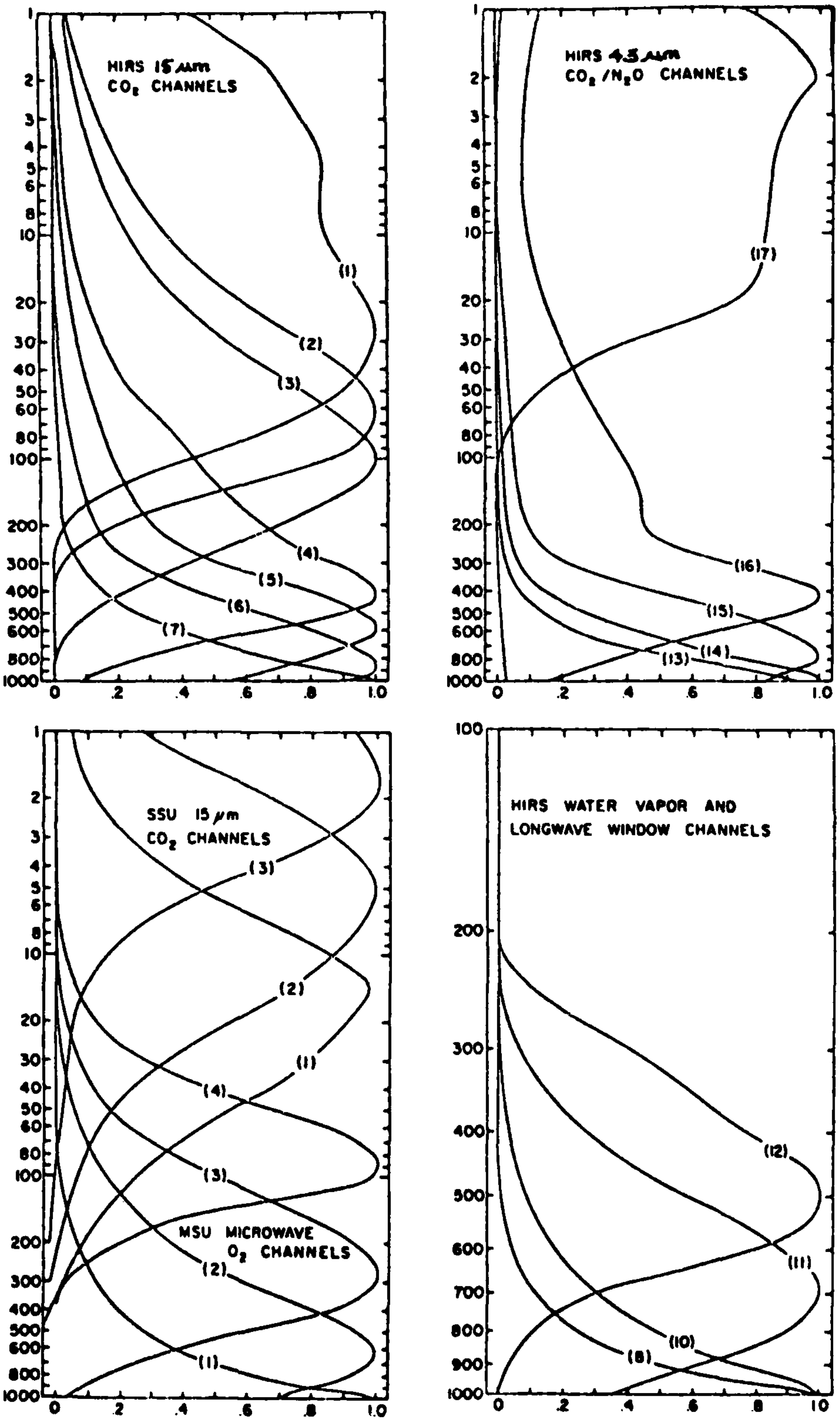

Fig. 4.1 TOVS weighting functions (from Smith et al., 1979).

4.4.2 Temperature Retrieval Assessment

A variety of recent studies have investigated the impact on NWP of satellite temperature retrieval (see Section 5.6). The ability to accurately resolve both vertical and horizontal temperature gradients at the scales required to delineate the atmosphere's three-dimensional temperature field is especially important for NWP. It has become apparent that the retrieval process itself may contribute significantly to the apparent impact (or lack thereof) of satellite temperature data (cf. Koehler et al., 1983). While the vertical resolution and absolute accuracy limitations of retrieved profiles are commonly cited deficiencies, their dependence on the retrieval method used is often neglected. However, objective assessment methods are available (Peckham and Flower, 1983). It is important to define the overall retrieval process in terms of its subelements including the instrument, the data preprocessing methods, and the retrieval algorithm (Halem et al., 1978; Phillips, 1983). Each of these components contributes directly and indirectly to the impact of the retrieved temperatures. Sources of error in these components include transmittance model errors, surface emissivity effects, instrument noise, vertical and horizontal sensor resolution, the effects of haze, clouds, precipitation, and absorption by non-uniformly mixed gases, the inability to correctly account for surface contributions to observed radiances, and smoothing or the neglect of physical mechanisms by the retrieval method.

4.4.2.1 Vertical Resolution

The vertical resolution of passive retrieved temperature profiles is fundamentally related to the pressure and temperature dependence of atmospheric transmission in the spectral intervals chosen for the measurement (Wark and Fleming, 1966; Chen et al., 1974). Vertical gradients of atmospheric transmittance (i.e. weighting function shape) are dependent on spectral region (e.g., IR vs. microwave), spectral resolution of the channel (ratio of channel width to center frequency, $\Delta\nu/\nu$), and the nature of the specific absorption features within the channel bandpass (i.e., line centers, line wings, etc.; see Staelin, 1983). Narrowing of the spectral intervals provides some opportunities for increase in vertical resolution, but decreases the signal to noise ratio. This tradeoff between instrumental noise and vertical resolution has been investigated (Conrath, 1972; see also Chahine, 1979) with the Backus-Gilbert formalism (see Section 4.3). A novel approach for enhanced vertical resolution exploiting both increased spectral resolution and the temperature dependence of high J-lines of the R branch of the 4.3 μm CO_2 band has been proposed for the Advanced Moisture and Temperature Sounder (AMTS) (Kaplan et al., 1977; Kaplan, 1977). Retrieval resolution is usually defined as the half width of the weighting function in local scale heights, which approximates the thickness of the layer contributing the primary atmospheric component to the integral term in Eq. (4-9). The half width of the weighting function, in turn, is roughly inversely proportional to the peak value of the unnormalized weighting function (Kaplan, 1977). Vertical resolution for the HIRS and MSU is approximately 5-6 km in the troposphere. Table 4.5 from Chahine (1983) summarizes characteristic weighting function widths for a variety of instruments. Vertical resolution is also dependent on the degree of smoothing introduced by the retrieval algorithm used (Rodgers, 1976; Thompson, 1982).

4.4.2.2 Layer Accuracy of Vertical Profiles

To provide some insight into the accuracy of temperature retrievals we discuss studies focusing on a single retrieval system, the TIROS N operational system (TOVS) described by Smith et al. (1979). While modifications are continually made to the operational system, the retrieval algorithm has changed little from the description given by Smith and Woolf (1976). Preprocessing of the data attempts to identify clear, partly cloudy and overcast soundings using a variety of techniques (Smith et al., 1979) and to discriminate between oceanic and continental soundings. Clear and partly cloudy retrievals use both infrared and microwave data while overcast retrievals use microwave data only.

A number of recent studies have evaluated the accuracy of TOVS retrievals. Phillips et al. (1979), Smith et al. (1980), and Gruber and Watkins (1982) compared collocated satellite retrievals and radiosondes. Schlatter (1981) and Koehler et al. (1983) spatially and temporally interpolated the NMC Limited-area Fine Mesh (LFM) analyses to the satellite sounding locations to provide data for verification. Fig. 4.2 provides a sample comparison of rms differences between radiosonde and satellite temperature soundings for one month (from Smith et al., 1980). These differences are commonly referred to as errors and we will follow that practice. They are of course only apparent

Table 4.5 Sounder resolutions. Relative spectral and vertical resolutions of several past, present, and future sounders (from Chahine, 1983).

Band	$\Delta\nu/\nu$	Half-width in scale heights	Remarks
Stratosphere			
14.5 μm	10^{-2}	2.4	VTPR/HIRS
15 μm	10^{-3}	1.6	AMTS
15 μm	10^{-4}	1.4	Wings of Lines
60 GHz	10^{-3}	1.3	AMSU
Troposphere			
15 μm	10^{-2}	1.6	VTPR
60 GHz	10^{-3}	1.5	MSU/AMSU
4.46 μm	10^{-2}	1.3	HIRS
4.18 μm	10^{-3}	0.69	AMTS
4.18 μm	10^{-4}	0.6	Wings of Lines

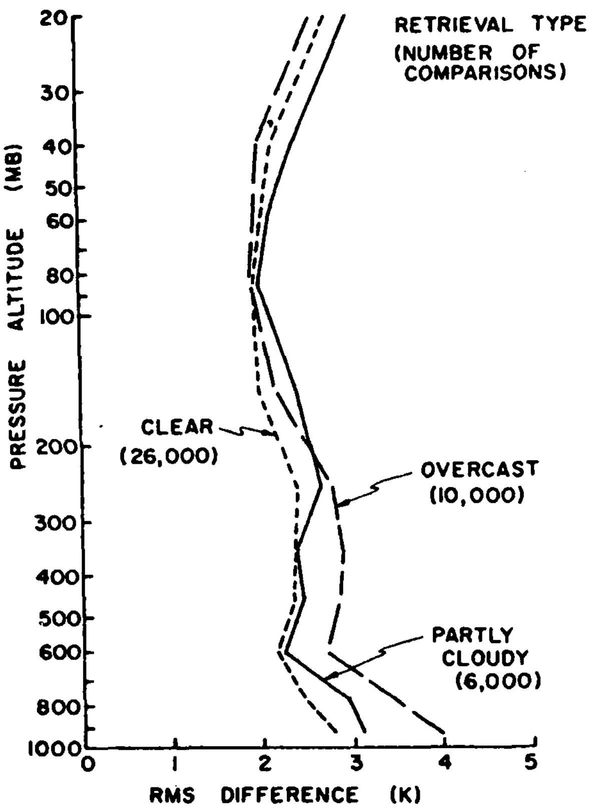

Fig. 4.2 Verification of satellite temperature soundings. Rms differences between radiosonde and satellite temperature soundings for April 1980 (from Smith et al., 1981a).

or perceived errors. In general, temperature accuracies throughout the middle and upper troposphere in clear and partially cloudy conditions are on the order of 2.25 K. The largest temperature errors generally occur near the surface and in the vicinity of the tropopause. Difficulties in the vicinity of the tropopause may be due to the vertical resolution properties of the instrument as well as the low radiation emanating from this layer. Errors near the surface may be due to the effect of surface emissivity, low level cloud, high terrain and/or other such factors. In the lower troposphere errors increase with increasing cloudiness, while above 200 mb overcast retrievals are better than clear retrievals. This is probably a consequence of preprocessing to identify partial cloudy and cloudy cases and reliance on microwave only sounding in overcast situations. Careful manual sounding selection using interactive computer techniques can improve the situation (Paulson and Horn, 1981). Slightly better performance of cloudy retrievals in the upper atmosphere may also be partially due to microwave only data (cf. Fig. 4.1). Biases are found in the retrieval of cloudy oceanic temperature profiles due to the predominance of continental soundings in the a priori data base (Phillips, 1981). This result is largely due to the nature of the statistical retrieval algorithm.

In order to isolate sources of error which are specific to either instrument or retrieval method, one can compare instruments and/or retrieval algorithms. GLAS, for example, has compared the use of physical retrievals based on the work of Chahine (Section 4.3) to the NESS operational algorithm (Susskind and Rosenberg, 1980; Susskind et al., 1982) (Fig. 4.3). There is a notable improvement using the physical retrieval (i.e., GLAS) methodology especially in the presence of clouds. This result conflicts with the recommendations leading to the adoption of statistically based retrievals at AFGWC (Mount et al., 1977).

It is expected that improved satellite sensors will lead to reduced temperature profile errors. This has been confirmed in the System 85 Test, a set of simulation experiments conducted cooperatively by NOAA and NASA (Phillips, 1983) which compare the current operational HIRS-2 and the proposed AMTS (Kaplan et al., 1977; Chahine, 1983) infrared sensors. This test also compared the operational statistical retrieval algorithm (STAT) and the GLAS physical retrieval algorithm (PHYS). Results for clear and cloudy cases are

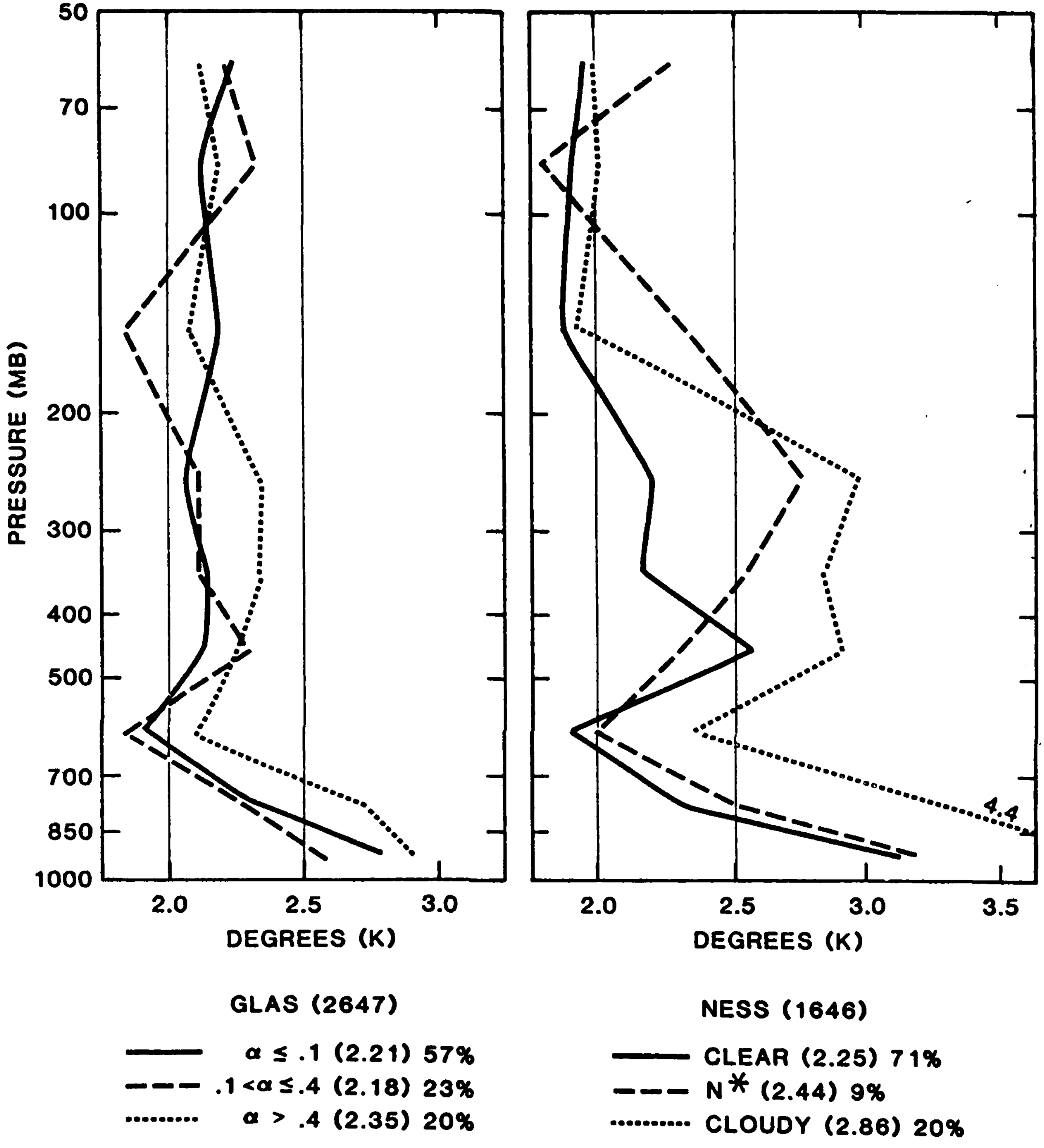

Fig. 4.3 Effect of cloud on retrieval error. Rms mean layer temperature errors compared to radiosondes within 3 h and 110 km versus degree of cloudiness for January 5 - January 15, 1979. GLAS retrievals are separated according to retrieved cloud fraction, α. NESS retrievals are separated according to reported retrieval type (from Susskind et al., 1982).

presented in Fig. 4.4 and Table 4.6 (from Phillips, 1983), respectively. Again, the physical retrieval method outperforms the statistical method.

4.4.2.3 Depiction of Synoptic Features

Because of their high spatial density, satellite soundings contain abundant information about the horizontal variation of temperature. The above accuracy analysis does not provide a complete view of the ability of satellites to depict details of the three-dimensional thermal structure of the atmosphere. Satellite-data derived temperature fields have generally been judged adequate to locate synoptic (Moyer et al, 1978) and mesoscale (Hillger and Vonder Haar, 1977; Mills and Hayden, 1983) features for semi-quantitative analyses. However, satellite derived thermal gradients, both vertical and horizontal, have been found to be weaker than those of a standard objective analysis (Moyer et al., 1978; Tracton et al., 1980, Schlatter, 1981; Gruber and Watkins, 1982; Koehler et al., 1983). In some cases, this may be directly associated with spatial smoothing due to the limited horizontal resolution of the sensor involved (e.g., Grody and Pellegrino, 1977). Koehler et al. (1983) point out that the reduced variance may be due to biases introduced because clouds are correlated with synoptic patterns and, within the operational retrieval system, cloudy retrievals are microwave only while clear retrievals use all channels. McMillan and Dean (1982) suggest that the regression-based operational retrieval algorithm tends to reduce the variance in the resultant temperature fields. Reduction of satellite observed temperature variance is a longstanding issue (Phillips et al., 1979; Tracton et al., 1980; Gruber and Watkins, 1982). Whether it is attributable to the instrument, cloud treatment or retrieval algorithms, the apparent correlation of satellite temperature errors with synoptic features (Koehler et al., 1983) has important implications for the assimilation process (Section 5.2).

4.5 Momentum (Winds)

Winds may be observed remotely from the motion of features, usually cloud features, seen in geostationary satellite imagery; from the pattern of sunglint over the ocean; from observed emissivities of the ocean surface; from radar backscatter from the ocean surface; and from lidar measurements of the motion of aerosols and water vapor features. The first three of these methods

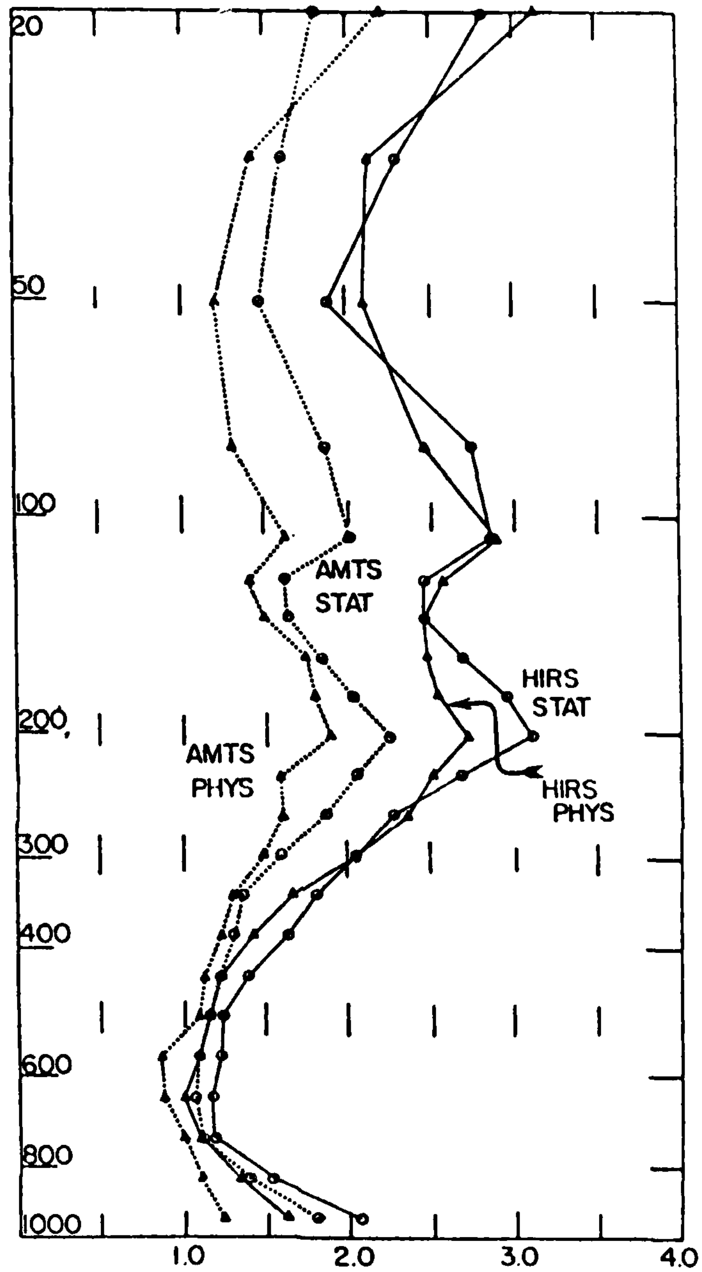

Fig. 4.4 System 85 clear column results. Vertical distribution of rms retrieval error averaged over all latitudes and both seasons. Values are plotted at the mean log p for each layer (from Phillips, 1983).

Table 4.6. System 85 cloudy results. Squared temperature errors obtained using statistical (STAT) and physical (PHYS) retrieval methods applied to simulated HIRS and AMTS radiance data in the presence of clouds. The numbers in parentheses are the true averaged variances of squared temperature in each layer for the clear and cloudy test profiles (from Phillips, 1983).

Sample Size		(Known) Clear Col 96	Lst Clds 13 or 12	Avg Clds 14	Most Clds 13 or 10	All Clds 40 or 36
		(53.00)	190-16 mb			(54.62)
HIRS	STAT	8.95	11.75*	10.63	9.02*	10.56*
HIRS	PHYS	9.14	9.01	8.53	6.32	7.97
AMTS	STAT	4.63	5.07*	4.15	3.61*	4.31*
AMTS	PHYS	3.33	3.40	3.77	2.88	3.36
Average		6.51	7.26	6.77	4.64	6.50
		(33.11)	464-190 mb			(33.15)
HIRS	STAT	9.24	9.65*	12.46	7.78*	10.22*
HIRS	PHYS	8.81	8.72	9.30	4.56	7.57
AMTS	STAT	6.18	7.05*	5.22	3.09*	5.24*
AMTS	PHYS	4.30	5.74	3.76	3.19	4.22
Average		7.13	7.77	7.69	4.55	6.76
		(43.45)	1000-464 mb			(34.97)
HIRS	STAT	2.78	2.91*	2.14	3.33*	2.73*
HIRS	PHYS	1.82	1.60	2.15	2.57	2.11
AMTS	STAT	2.37	1.36*	1.73	1.95*	1.67*
AMTS	PHYS	1.47	0.85	1.02	1.42	1.09
Average		2.11	1.66	1.76	2.28	2.12

* The STAT method omitted 1 retrieval in the "LEAST CLD" sample and 3 in the "MOST CLDS" sample, for both instruments. None of these 4 cases represented extreme errors in the PHYS retrievals, however.

are passive and the last two are active. These measurements all provide single level winds, except for lidar winds. Although there are several methods to observe surface winds remotely, it is difficult to assimilate these observations because the model calculates its own surface wind at every time step from the wind at its lowest levels (Section 3.3.3). We discuss the assimilation of surface winds in section 5.6.

Cloud track winds (CTWs) are obtained by identifying distinctive features in successive cloud images. Usually this is done by a human aided by a McIDAS system (Hubert, 1976; Mosher, 1979; Endlich and Wolf, 1981). Typically these winds are assigned only vague heights, such as "high" or "low," or arbitrarily assigned to some pressure level. If two geosynchronous satellites observe the same cloud, heights may be obtained geometrically (Hasler, 1981). Alternatively, proper height assignment is possible if, as subsequently described (Section 4.6.1), cloud top temperatures can be obtained and compared to a known temperature profile (Mosher, 1976; Reynolds and Vonder Haar, 1977). Even so, it is not clear whether the motion vector should be assigned to the cloud top (Lee, 1979). Hasler et al. (1976) find that CTWs for tropical cumuli have the best agreement with in situ winds at cloud base. The imagery used is often in the visible region of the spectrum, but these techniques may be applied in any spectral region in which distinctive features may be identified. Menzel et al. (1983) tracked features in several VAS channels of the 14 μm CO_2 band thereby obtaining CTWs at several levels simultaneously. Water vapor features in the 6.7 μm region have also been used to examine the dynamics of mid- to upper-tropospheric moisture and a means to track winds at these levels (Allison et al., 1972; Steranka et al., 1973; Rodgers et al., 1976; Mosher and Stewart, 1981; Fischer et al., 1981; Eigenwillig and Fischer, 1982). Typical accuracies obtained from CTWs are shown in Table 4.1. CTWs are part of the FGGE Level II data sets but reassignment of at least the Japanese CTWs is necessary (Baker et al., 1981).

Variations in sunglint patterns over the ocean due to changes in surface wind speed have long been observed in polar orbiter visible imagery (Kornfield, 1974; Wald and Monget, 1983; and references therein). Studies using NOAA-6 AVHRR data, for example, have indicated good qualitative agreement between conventional surface winds and those determined from sunglint (Takashima and Takayawa, 1980). Wylie et al. (1981) present a comparison of sunglint microwave, and cloud-track winds. Due to temporal and spatial limitations, however (Fett and Mitchell, 1977; Wylie et al., 1981), the operational use of sunglint winds is limited.

Microwave ocean surface emissivity has been measured by a variety of sensors and depends on wind speed as well as other parameters (Wentz, 1983). Algorithms for the retrieval of surface wind speed from microwave radiances have been developed and applied (Stogryn, 1967; Hollinger, 1970; Ross and Cardone, 1974; Webster et al., 1976; Wilheit, 1979; Wilheit and Chang, 1980). Ocean surface wind speed will be available from the DMSP SSM/I microwave imager. Simulations suggest that except in high precipitation (~5 mm/hr) conditions, wind speed will be retrieved with an rms error of 1.2 m s^{-1}.

Unique microwave data sets were available from the SEASAT mission. SEASAT carried three surface wind sensors - SMMR (a radiometer), SASS (a scatterometer), and an altimeter. The last two instruments actively measure ocean surface properties related to the surface wind using microwave radars. The altimeter measures significant wave height which may be related to wind speed (Brown, 1979; Brown et al. 1981; Chelton et al., 1981). The scatterometer measures backscatter from capillary waves which are forced by the surface wind (Jones et al., 1977). For SMMR and the altimeter only wind speed may be estimated, but the scatterometer is sensitive to the orientation of the capillary waves and thus to wind direction, thereby providing surface wind vectors (Jones et al., 1977; Pierson et al., 1980). Unfortunately, SASS does not determine the wind vector uniquely; partial solutions to this ambiguity problem are available (Barrack et al., 1980; Hoffman, 1982) but the most satisfactory solution apparently is an additional antenna in the design of the next scatterometer. All the SEASAT winds compare favorably with each other and with high quality surface winds (Wurtele et al., 1982; Brown et al., 1982; Wentz et al., 1982). Different comparisons gave rms wind magnitude differences of order 1-2 m/s. SASS wind directions attained the design goal accuracy of ±20°. SEASAT data is available from EDIS but only for the three month mission lifetime July through September 1978.

Doppler lidar from space is at present only a proposal (Lawrence et al., 1982) but its promise is great in terms of accuracy and vertical resolution. Lidar wind measurements require the presence of aerosol and are contaminated by clouds. CTW techniques can be applied to a sequence of aerosol patterns observed by lidar (Atlas and Korb, 1981; Sasano et al., 1982) or winds can be inferred from the Doppler shift measured by lidar (DeMarzio et al., 1979; Schwiesow et al., 1981).

4.6 Moisture and Cloudiness

Radiatively, water is the most important constituent of the troposphere. Clouds and water vapor profoundly affect the temperature retrieval problem and the parameterization of radiative heating in the atmosphere (Section 3.2). In addition, the variability of clouds and water is substantial, occurring on a variety of space and time scales (Section 5.2). A variety of studies have found that updating the moisture field can improve short-range precipitation forecasts (Section 5.2), and that latent heat release associated with clouds can have large, short-term influences on storm development (Section 3.6). Satellite retrievals of moisture and clouds have not yet been routinely used in numerical weather prediction, despite the ready availability of cloud imagery and retrieved integrated water vapor, probably because of the difficulty of translation into the forecast model requirements of three-dimensional fields of clouds and of water vapor mixing ratio. Some studies which use available data retrieved from satellites to analyze humidity are discussed in Sections 5.2 and 5.4. In this section we discuss the more detailed remote sensing of moisture (i.e., water vapor) distribution (4.6.1), cloud parameters (4.6.2), and precipitation (4.6.3).

4.6.1 Water Vapor

Absorption features used for passive remote sensing of water vapor include the infrared rotation-vibration bands near 6.7 μm, the infrared pure rotation bands near 20 μm and the microwave rotational lines at 22.235 and 183.310 GHz. A comprehensive bibliography of moisture retrieval related literature has recently been prepared (Rosenberg et al., 1983b). Application of Eq. (4-8) to the forward problem for water vapor indicates sensitivity to both surface emission and temperature profile. Thus the water vapor problem is directly affected by uncertainties in the coincident temperature retrieval. Problems associated with water vapor profile retrieval include (Conrath, 1969; Chahine, 1972; Smith and Woolf, 1976; Hayden et al., 1981): (1) the inability to resolve vertical structure due to broadness of weighting functions; (2) sensitivity of retrieved humidity structure in the lower atmosphere to errors in surface emission; (3) errors due to the non-linear dependence of the humidity retrieval on the temperature retrieval; and (4) errors due to the dependence of water vapor weighting functions on the water vapor profile.

As a practical consequence of this last point, it is difficult to select a set of sensor channels a priori which optimally span the vertical domain. Figure 4.5 illustrates the weighting functions for the TIROS-N HIRS-2 water vapor channels and their dependence on water vapor abundance. Note that an increase in water vapor results in a decreased signal, since the weighting functions are shifted to greater heights where $B_i[T(p)]$ in Eq. (4-8) is smaller.

Due to the lack of success in obtaining operationally useful water vapor profiles, much research has focused on column integrated (i.e., total precipitable) water vapor using infrared (Shen and Smith, 1973; Prabhakara et al., 1979) and microwave (Staelin et. al., 1976; Grody and Pellegrino, 1977; Liou and Duff, 1979; Chang and Wilheit, 1979; Grody et al., 1980; Chelton et al., 1981; Prabhakara et al., 1982; Alishouse, 1983) sensors. The microwave resonance at 22.235 GHz is rather weak (Gaut, 1968; Waters, 1976) and, therefore, reliable sounding throughout the atmosphere can not be achievable. The microwave studies are limited to cases over the ocean since land emissivity is quite variable at microwave frequencies. Accuracies reported for the retrieval of precipitable water range from 0.2 to 0.45 g cm^{-2} rms error (Staelin et al., 1976; Haydu and Krishnamurti, 1981; Prabhakara et al., 1982). Precipitable water determinations are expected from DMSP polar orbitors when the channel SSM/I microwave imager (Hughes Aircraft Company, 1980) becomes operational. In simulations, the rms error in precipitable water has been estimated as 0.64-0.80 g cm^{-2} over land, depending on season, and 0.20 g cm^{-2} over the oceans. Simulated retrievals of precipitable water using a previous similar instrument (SSH) and a statistical retrieval method (Gaut et al., 1972) yielded rms errors of 0.6 gm/cm^2 (Crane, 1976).

Water vapor profile data from DMSP is available from the SSH/2 infrared sounder (Barnes Engineering Company, 1978) containing seven broadband channels in the 18-28 μm rotational band and one centered at 12.5 μm. (After flight of the remaining SSH/2 instrument no further DMSP IR sounders are planned.) A recent study of the SSH/2 capabilities for moisture profiling (Rosenberg et al., 1983a) concluded that retrievals were not possible down to the surface layers and that maximum sensitivity to moisture was in the middle troposphere

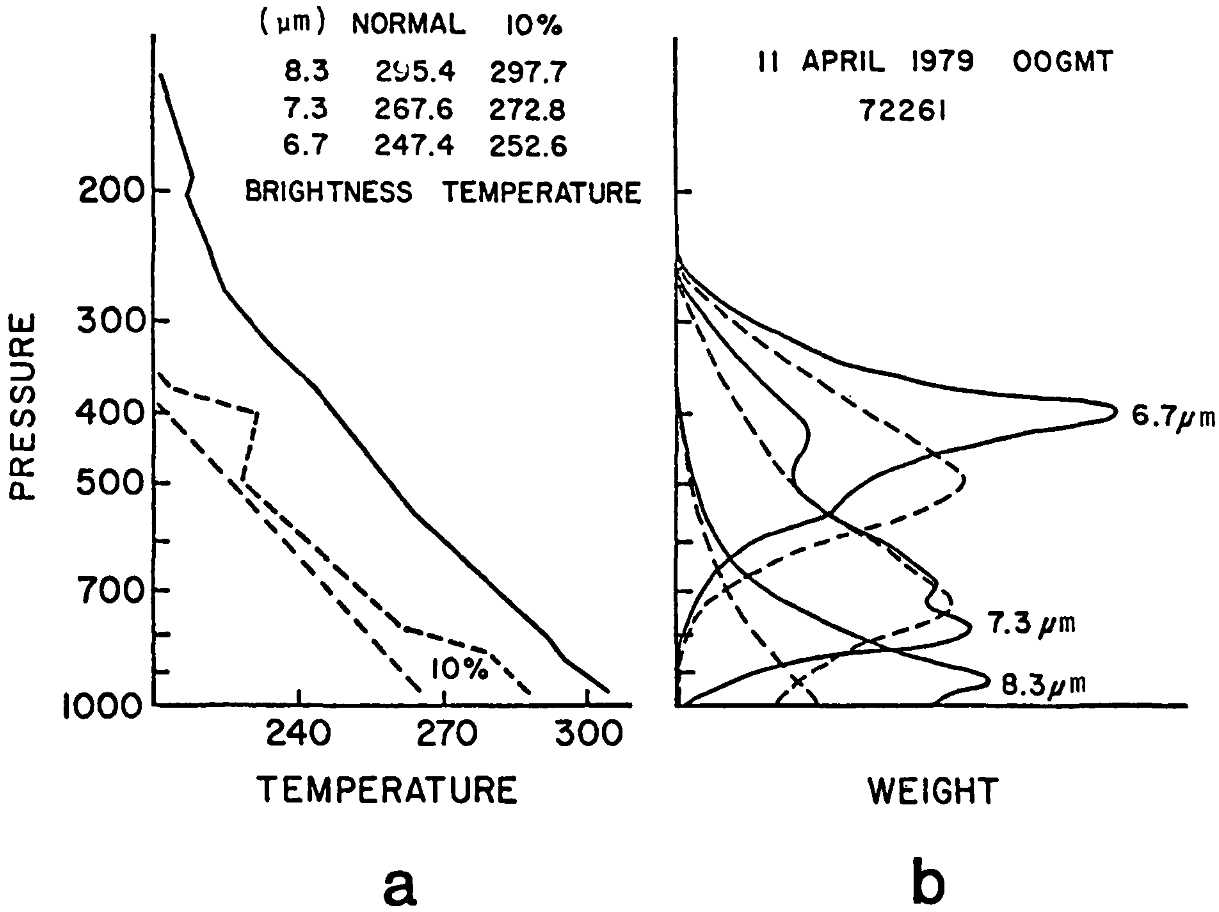

Fig. 4.5 TOVS water vapor weighting functions. (a) Midlatitude standard atmosphere temperature and dewpoint profile (10% humidity profile also shown). (b) Brightness temperature weighting functions for TOVS water vapor channels for profiles shown at left (from Hayden, 1980).

(peaking from 400-600 mb for a mid-latitude atmosphere). Figure 4.6 based on calculations at AER tends to support these conclusions, but use of the 12.5 μm "window" channel would help retrieve lower level water vapor, especially over open ocean, as it is sensitive to boundary layer water. The temperature channels are also sensitive to water vapor in the lower troposphere.

As yet there is no operational microwave water vapor profiler. Use of the strong 183 GHz absorption feature has been suggested (Schaerer and Wilheit, 1979; Rosenkranz et al., 1982; Kakar, 1983), however, and an aircraft instrument implementation called the Advanced Microwave Moisture Sounder (AMMS) has been flown (Wang et al., 1983). Comparisons of relative humidity profiles retrieved at three levels (using three frequencies) with collocated radiosondes are promising, but surface emissivity effects over land are still a serious problem.

Given the inability to accurately prescribe a vertical profile, operational products are currently limited to moisture values (precipitable water) for three layers: surface - 700 mb, 700-500 mb, and above 500 mb (Smith, 1980). Prescribed accuracy goals for these retrievals are ±30% (NOAA, 1981). In spite of these limitations, the ability of remotely sensed data to delineate the horizontal structure of relevant water vapor features has been demonstrated (Hillger and van der Haar, 1977, 1981; Moyer et al., 1978; Smith and Zhou, 1982; Chesters et al., 1983). Both high horizontal resolution of the moisture data and improved moisture determination through the application of processing techniques beyond those used operationally have been cited as positively impacting the depiction of moisture structure with promising implications for NWP (Hayden et al., 1981; Mills and Hayden, 1983). The full potential of the use of satellite moisture data in NWP models will not be realized, however, until more effective algorithms have been developed for retrieving vertical structure from satellite flux measurements and other ancillary information.

4.6.2 Cloudiness

Among the first remotely sensed data pertaining to synoptic scale weather were photographs of extended cloud systems obtained from high altitude balloons and rockets (Greenfield, 1982). The potential application of these observations to meteorological analyses was quickly recognized during the

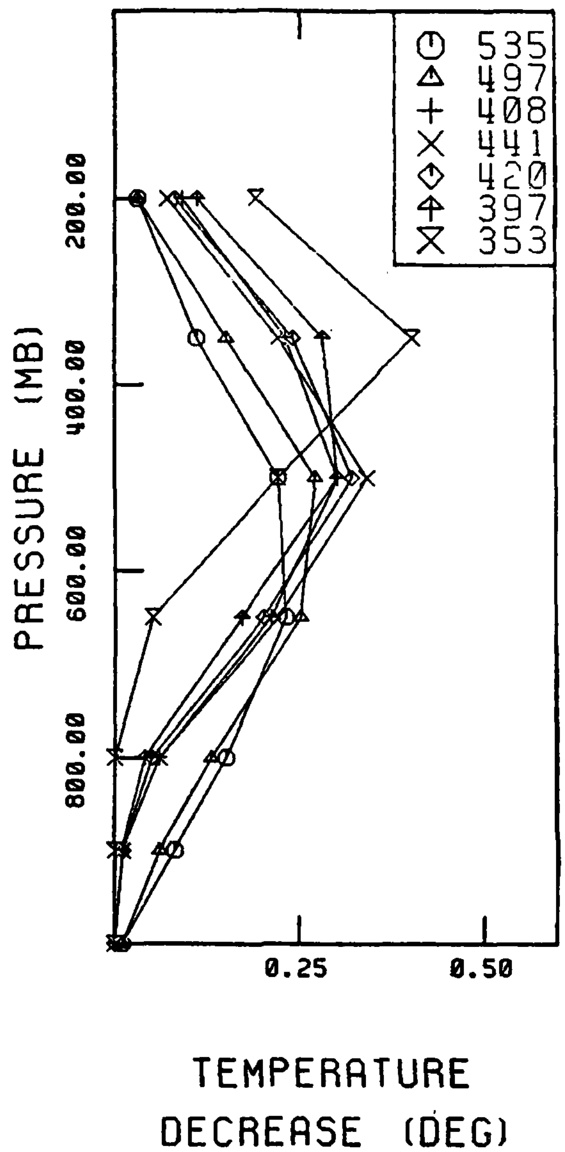

Fig. 4.6 DMSP SSH/2 water vapor sensitivity. Sensitivity at pressure p is defined as the brightness temperature decrease relative to a standard case for a 20% increase in water vapor at pressure p with all other variables fixed at their standard values. Each channel (in cm^{-1}) is represented by a particular symbol as indicated in the box at the upper right.

early TIROS series of satellites (Tepper, 1982) and the use of satellite cloud imagery has subsequently been a major influence on forecasting and analysis methods (Conover, 1962, 1963; Anderson et al., 1969; Oliver and Bittner, 1970; Fett and Mitchell, 1977). Imagery is of special interest to Southern Hemisphere meteorologists (Section 2.3). These early techniques relied primarily on subjective and qualitative evaluation of cloud morphology. Recently, the importance of clouds in radiative processes (cf. Ohring and Clapp, 1980; Wang et al., 1981; Ohring et al., 1981) has focused attention on objective techniques to obtain quantitative cloud parameters from remotely sensed measurements for operational (Fye, 1978) and research oriented (WMO, 1982; Schiffer and Rossow, 1983) purposes.

This section reviews a variety of cloud analysis techniques which exploit available satellite imager and sounder data to infer cloud parameters. In general, cloud parameters define the vertical location, horizontal distribution, and optical properties of the clouds and thus provide information on sources and sinks of thermal forcing mechanisms related to clouds within the atmosphere (Chahine, 1982). Cloud parameters include: (1) bulk parameters such as total cloud cover (F) and cloud cover fractions of high, middle, and low clouds ($F^{h,m,l}$, respectively), corresponding cloud top heights ($Z^{h,m,l}$) and thicknesses ($\Delta Z^{h,m,l}$); (2) optical properties such as visible optical depth (τ) and albedo (A) and infrared emissivity (ϵ); and (3) information related to cloud microphysics such as liquid water content, cloud type, cloud phase (i.e., ice/water), and possibly dominant particle size. Cloud parameters potentially retrievable from satellite data are summarized in Table 4.7. Not all of these quantities are independently available as supplementary data may be required. For example, if cloud top temperature and cloud top pressure or height are obtained they will be related by the clear column atmospheric temperature profile. Supplementary data which may be required include the surface emissivity ϵ_s, surface temperature T_s, surface albedo A_s, and water vapor (or relative humidity) profile.

With the exception of the threshold approaches for visible and infrared imager data processing which are currently used in the Air Force 3DNEPH (Fye, 1978), the approaches described below have not been implemented operationally. Also, none have been used to provide operational initial data for NWP models.

Table 4.7 Retrievable cloud parameters. These cloud parameters are potentially retrievable from satellite data.

Variable	Definition
F	cloud coverage fraction
T_c	cloud top temperature
Z_c	cloud top height
ΔZ_c	cloud thickness
ε_c	cloud emissivity
τ_c	cloud optical depth
A_c	cloud albedo
N_c	cloud particle number density
M_c	cloud liquid water/ice content
CT	cloud type
CP	cloud phase (ice/water)
CL	cloud layer information

4.6.2.1 Analysis Approaches

Extant cloud parameter analysis techniques may be categorized based on the type of satellite data analyzed, the required auxiliary data sources, the computational level of effort, or the mathematical formalism. Alternatively, classification according to the rough physical basis of each technique produces four general methods: threshold, statistical, bispectral, and sounder-based (Table 4.8). Previous reviews of these methods (Smith, 1981; WMO, 1982) have summarized the capabilities of the various techniques vis-a-vis the needs of the climate modeling community and have generally emphasized approaches used in conjunction with imagery, i.e., threshold and statistical methods. The requirements for NWP of vertical temperature profiles makes attractive those techniques which provide cloud parameter information as a by-product of the general retrieval process for vertical temperature and moisture profiles.

Threshold methods are rooted in the observations of early satellite imagery analysts that in visible imagery clouds generally appear brighter and in infrared imagery clouds generally appear colder than the background (Anderson and Smith, 1970, Chapter 6). In applying this simple viewpoint to automated analyses, fractional cloud cover is assumed proportional to the difference between satellite measured radiance and a specified *threshold value of radiance*, which supposedly characterizes cloud-free conditions. This method is the basis of the visible and infrared satellite data processors of the AFGWC automated cloud analysis model (Fye, 1978). The advantage of the threshold method is its computational simplicity. Simplest single channel, fixed threshold algorithms, e.g., that of Saunders and Hunt (1981) require only two arithmetic operations per pixel. Two channel techniques calculate coverage fraction from visible data and cloud top altitude by relating infrared-based equivalent blackbody temperature (EBBT) to height (Ackerman and Cox, 1980; Chen et al., 1981). A shortcoming of using fixed thresholds to characterize cloud-free conditions is the relative non-uniformity of visible and infrared radiances reflected and emitted respectively, by the Earth's surface. Variable thresholds based on histograms of the spatial variation of radiance (e.g., Fye, 1978) or on extremum values of radiance climatologies may be used to overcome this problem. Even so, the data processing system will be tricked sometimes by snow cover and sunglint in the visible and by high terrain and temperature inversions in the infrared. Operational experience and auxiliary

Table 4.8 Summary of cloud parameter analysis methods. Numbers in parenthesis indicate spectral region. Cloud parameters are defined in Table 4.7. Subscript s indicates a surface quantity, T(Z) and RH(Z) are temperature and relative humidity profiles.

Method	Technique	Satellite Data	Supplementary Data	Cloud Parameter(s) Inferred
Threshold	1 channel	VIS	-	F
		IR(11)	T(Z)	F, T_c, Z_c
	2 channel	VIS, IR(11)	T(Z)	F, T_c, Z_c
Statistical	Regression	VIS	-	F, A_c
		IR(11)	T_s	F, ΔZ_c
	Graphical 1-D Histogram	VIS	-	F
		IR(11)	-	F, T_c
	2-D Histogram	VIS/IR(11)	T(Z)	F, T_c, Z_c, CT, CL
		IR(11)/WV(6)	-	T_c, Z_c, CT, CL
	3-D Histogram	VIS/IR/WV	-	F, T_c, Z_c, CT, CL
	Spatial Coherence	IR(11)	T(Z)	F, T_c, Z_c
		IR(11)/IR(3)	T(Z)	F, T_c, Z_c, CL
Bispectral		VIS/IR(11)	T(Z), ε_s, T_s, A_s, ε_c, A_c	F, T_c, Z_c
		VIS/NIR	-	F, CP
Sounder	2-channel IR	IR($2CO_2$)	-	F, T_c, Z_c
	2-channel MW			M_c
	6-channel IR/MW	IR(11)/MV(6)/ IR($2CO_2$)/MW(2)	T(Z), RH(Z)	F, T_c, Z_c, ΔZ_c, M_c, CL
	Multispectral		-	F, T_c, Z_c, ΔZ_c, ε_c, τ_c, A_c, N_c, M_c, CL

data (e.g., snow cover and terrain heights) help to alleviate these difficulties. Another problem is that deviations from assumed cloud optical properties (e.g., visible optical depth less than infinity and infrared emissivity less than one) can be confounded with cloud coverage changes. The method is therefore not applicable when cirrus is present.

A variety of analysis techniques employ statistical methods. For example, the threshold method may be augmented by using a histogram to chose an appropriate threshold value. Here statistical methods are defined as those which rely primarily on the statistical properties of satellite data. They include regression, graphical, histogram, spatial coherence, and dynamic techniques.

Statistical regression may be used to provide a semi-empirical predictor equation for selected cloud parameters using available satellite sensor data. Examples include univariate correlation of low level cloud thickness obtained from conventional observations with visible radiometer brightness levels (Kaveney et al., 1977), correlation of surface observed cloud amount with GOES visual imagery albedo (Keegan and Niedzielski, 1981), and correlation of 3DNEPH total cloud data amount with monthly mean albedo derived from a visible scanning radiometer (Hughes and Henderson-Sellers, 1983). While such relationships are simple to implement, they are difficult to interpret physically, they may be biased if the data is not representative, and there may be satellite sensor specific biases depending on other factors (e.g., sun angle, bandpass, etc.). For example, the study by Keegan and Niedzielski (1981) indicated a site specific and seasonal trend in the regression relations. While multivariate analyses may account for the effect of other factors difficulties in interpretation and data sampling remain.

Practical application of fixed threshold methods has motivated examination of the frequency distribution of radiance values over areas containing many (e.g., 64 for the 3DNEPH) individual resolution elements. Frequency distributions are expected to peak at radiance values corresponding to completely cloudy and cloud-free conditions. Techniques which use this idea to pick a threshold, i.e. cloud-free value of radiance, are termed graphical histogram techniques. For example, Smith et al. (1970) applied a one-dimensional histogram technique to the determination of cloud free fields-of-view using high resolution infrared scanner data. Histograms are routinely used to determine cloud/no cloud thresholds in the 3DNEPH infrared satellite data processor (cf.

Fye, 1978, Fig. 12). Extensions to two- or three-dimensional histograms (Shenk et al., 1976; Raschke et al., 1981) generally use combinations of visible and infrared data. Each cloud type is associated with a rectangular volume in the coordinate system defined by the observed radiances. Then, coverage fraction for each cloud type is evaluated from the visible data while infrared data (and auxiliary temperature profile information) are used to provide cloud top height. Characteristically, frequency contours are very irregular when displayed using multidimensional histogram techniques. Therefore clustering methods have been suggested as an alternative to the fixed box approach. Clustering methods (Desbois et al., 1982) have the advantage of forming natural groupings rather than relying on a priori classification; however, they require considerable computational effort.

The spatial coherence approach, like the histogram approach, uses the local structure of the infrared radiance field to determine radiances associated with cloud-free and completely cloudy fields of view. This technique is based on the recognition of characteristic patterns in plots of the local standard deviation versus the local mean of the brightness temperature of a single channel. For a single cloud layer, the characteristic pattern is an arch (Fig. 4.7) since in fields of view containing cloud boundaries the radiance is highly variable. Consequently fields of view which are partially cloudy may be identified (Coakley and Bretherton, 1982). For single-layered systems, the cloud coverage fraction, cloud top temperature, and perhaps ground temperature may be obtained; the technique allows for partial beam filling. Additionally, the method identifies multilayered clouds, subresolution clouds, and variable emissivity clouds. Treatment of multi-layered cloud systems using two channels provides the nonoverlapped fractional cover contributed by each layer (Coakley, 1983). Effective cloud top heights of each layer may be deduced from the infrared EBBT if auxiliary temperature profile information is available. Recently this approach has been extended to address the problems caused by cirrus and subresolution clouds (Molnar, 1983). As with other methods using imager data, variations in surface background may be mistaken for cloud fraction changes, but this technique, which may be automated, is particularly promising, especially over oceans.

Cloud parameters vary in time as well as in space. In general, the evolution of cloud parameters could be investigated by a time series using one of

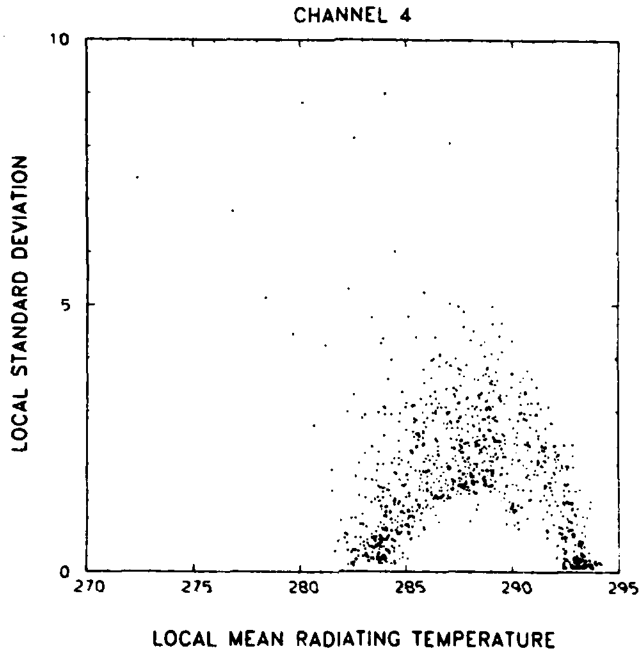

Fig. 4.7 The spatial coherence arch. An example of the 11-μm local mean brightness temperature and local standard deviations for 8x8 arrays of AVHRR data points. The cluster of points near 293 K represents cloud-free scan spots; the cluster near 283.5 K represents cloud-covered scan spots. The points between these clusters represent partially covered fields of view (from Coakley and Bretherton, 1982).

the above techniques. For example, Szejwach and Desbois (1979) have studied the temporal and spatial evolution of cloud patterns by applying cumulative histogram techniques to geostationary infrared imager data. For longer time scales polar orbiter data may be acceptable (Cahalan et al., 1981).

The bispectral method proposed by Reynolds and Vonder Haar (1977) provides a physical basis to the treatment of two channel visible-infrared sensor data currently available from a variety of meteorological imagers (e.g., GOES/VISSR, TIROS-N/AVHRR, DMSP/OLS). The calibrated visible, R_V, and infrared, R_I, radiances (only the latter require absolute calibration) may be simply expressed as functions of cloud coverage fraction F and cloud top EBBT, T_c, as:

$$R_V = [(1-F)\ A_s + FA_c]\ S_o \tag{4-10a}$$

$$R_I = (1-F)\ B\ (T_s) + F[(1-\varepsilon_c)\ B\ (T_s) + \varepsilon_c\ B\ (T_c)] \tag{4-10b}$$

where we have assumed a single cloud layer and where S_o is incident solar irradiance, A_s and A_c are surface and cloud bidirectional reflectances, respectively, $B(T_s)$ is surface infrared emission and ε_c is cloud emissivity. If these variables are known, (4-10) may be solved for cloud amount and cloud top temperature, and cloud top height, Z_c, may be deduced from an independently specified temperature profile. The bispectral method is attractive because of its physical basis (Eq. (4-10)) and its capability to correct for global variations of surface properties and varying illumination conditions caused by the satellite sensor/sun geometric configuration. However, it is restricted to daytime only and is not able to treat optically-thin cirrus. With respect to cirrus clouds, Platt (1983) has compared observational bispectral histograms with the relationship between R_V and R_I expected for idealized cloud layers. These results suggest that unbroken cloud layers with variable optical depth (i.e., cirrus) may be distinguished from broken clouds with uniform high optical depth (i.e., stratocumulus).

An extension of the bispectral analysis scheme (Rossow et al., 1980) is based on comparison of observed calibrated visible and infrared scanner radiances to extensive tables of theoretical radiances. The radiances are precalculated assuming either completely clear or completely overcast fields of view and using a complete radiative transfer treatment. The visible radiance calculation includes variable viewing geometry, surface reflectance and cloud optical depth and a complete multiple scattering calculation, which assumes plane parallel clouds and a specified particle size distribution. Infrared radiances are calculated as functions of viewing geometry and cloud top height for model atmospheric absorber profiles of water vapor, etc. The method determines cloud visible optical thickness, τ_c, cloud top height, Z_c, cloud top EBBT, T_c, cloud coverage fraction, F, and infrared emissivity, ϵ_c. Determination of infrared emissivity from visible optical depth allows more realistic determination of cirrus cloud top height and temperature. The calculations required to implement this method are considerable, but are performed off-line.

Multispectral/sounder analysis techniques apply the principles of radiative transfer theory to the analysis of cloud parameters using absolutely calibrated radiances available from satellite sensors designed to retrieve vertical profiles of atmospheric temperature and moisture. Since cloud seriously affects these retrievals, methods were sought to identify cloud free and partially cloudy fields of view (Smith, 1967; Chahine, 1970). Thus, the ability to extract cloud properties emerges as a by-product of the temperature profile retrieval process (Chahine, 1982). Conceptually, these methods are simple and require the fewest a priori assumptions regarding cloud radiative properties and spatial scales (Wielicki and Coakley, 1981). For example, cloud fractional coverage within an instrumental field of view and cloud emissivity may be less than unity. In practice, however, the computational effort required is proportional to the number of cloud parameters desired. Comparisons of results with retrieved quantities have largely been limited to cloud heights and coverage fractions (McCleese and Wilson, 1976; Chahine et al., 1977; Smith and Platt, 1978; McCleese, 1978).

The fundamental assumption involved is that the horizontal spatial variation of clouds is greater on smaller scales than that of other atmospheric variables, in particular temperature. It is assumed that radiance differences between adjacent fields of view are largely due to differences in cloud. That is, the clear column radiances in adjacent fields of view are assumed to be the same. Then, radiance data measured in two spectral regions from two adjacent fields of view may be used to obtain clear column radiance for use in obtaining the temperature retrieval and to obtain cloud amount and height within each field of view (Chahine, 1970, 1974, 1975, 1977a; Smith et al., 1970; Bunting and Conover, 1974). The first experimental verification of this method is reported by Shaw et al. (1970). A comprehensive error analysis of these methods has been performed by Wielicki and Coakley (1981). Results indicate that errors in cloud top pressure and amount depend on cloud height and amount within the field of view. In general, accuracy is best for middle and high level clouds (above 3-4 km). In contrast large errors are found for low level clouds presumably due to surface effects. Additionally, cloud retrieval errors caused by uncertainties in clear sky radiance can be large.

While cloud amount and pressure based on a single layer cloud model may be inferred on the basis of just two sounding channels, there are advantages to a multispectral approach. These include the ability to infer a larger set of cloud parameters and increase of cloud "signal" over observing system noise. Chahine (1982), has presented a technique based on nonlinear minimization methods which is applicable to a general set of sounding channels. This approach minimizes the errors in cloud parameter determination caused by uncertainties in clear column profile characteristics. For example, microwave channels may be exploited to infer clear column temperature and moisture profiles.

Parameterized and hence less general multispectral approaches have been applied to investigate the effect of multilayered clouds including cirrus on high resolution infrared sounder channels (Feddes and Liou, 1977a, b; Liou et al., 1977; Liou et al., 1978; Feddes and Liou, 1978) and combined infrared-microwave systems (Liou et al., 1980b, c; Yeh and Liou, 1983). The primary advantage of microwave frequencies within this context is that cirrus clouds are transparent so that only water clouds are sensed. This allows cloud liquid water content of low cloud over the ocean to be inferred (Liou and Duff, 1979) and partially removes the cirrus and low and middle cloud ambiguity (Yeh and Liou, 1983). Chahine (1982) noted experimentally that effective infrared cloud amount in regions where low clouds are obscured is less when determined from infrared data than that determined from visible imager data largely because cloud infrared emissivity is less than unity (Yamamoto et al., 1970). Note, however, that cloud emissivity could be estimated from microwave liquid water content measurements using the parameterization of Chýlek and Ramaswamy (1982).

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While multispectral methods provide the potential for the most complete retrieval of cloud parameters they require considerable computer resources. However, this disadvantage is of no account when such methods are implemented within an operational temperature retrieval system which must remove the effect of cloud anyway.

4.6.2.2 Cirrus Cloud

Most of the cloud parameter analysis methods described above fail when cirrus cloud is present (Smith, 1981; WMO, 1982). This is a serious deficiency considering the extent of cirrus coverage and the role of cirrus in the global radiation budget (Platt, 1983). There are also important military requirements for the prediction of cirrus cloud coverage, particle density, and thickness (Conover and Bunting, 1977; Hardy, 1979). Remote sensing of cirrus cloud optical properties is limited by uncertainties in the characterization of their microphysical properties (Bunting, 1980). Deficiencies in our understanding of cirrus are gradually being corrected by the combined results of studies both theoretical (Hansen, 1969; Liou, 1972, 1973, 1974, 1977; Liou and Stoffel, 1976; Feddes and Liou, 1977; Stephens, 1980) and observational (Murcray et al., 1974; Heymsfield, 1977; Platt et al., 1980; Paltridge and Platt, 1981; Platt and Dilley, 1981; Heymsfield and Platt, 1983).

Since cirrus clouds are semitransparent at infrared wavelengths (Valovcin, 1968), their contribution to satellite incident radiance is a function of three unknowns: cloud emissivity (ε_c), coverage fraction, and blackbody radiance from below (see Equation 4.10b). Threshold methods are not applicable since they can only infer a single cloud parameter. Bispectral methods provide two channels of information usually used to infer cloud coverage fraction and cloud top temperature assuming a fixed emissivity. However, cirrus clouds have variable emissivity. A relationship between visible reflectivity and infrared emissivity may be used to eliminate one unknown (Shenk and Curran, 1973; Reynolds and vonder Haar, 1977). Platt (1983) has demonstrated that cloud amount, albedo, and emissivity may be obtained within the context of the bispectral method from the patterns of two-dimensional histograms. The cases of broken (low and middle) cloud with uniform high optical depth and unbroken (cirrus) cloud with variable optical depth are distinguished by their characteristic bispectral frequency distribution patterns.

In order to avoid errors introduced by specifying a relationship between visible albedo and infrared emissivity, Szejwach (1982) has developed a technique using two infrared channels aboard METEOSAT; the 10.5 - 12.5 μm window channel and the 5.7 - 7.1 μm water vapor channel. The technique determines actual cloud top emission temperature by exploiting the observation that the effective emissivity of cirrus cloud is almost perfectly correlated in the two channels, thus reducing the number of unknowns by one. Once emission temperature is determined either channel radiance provides the effective emissivity (i.e., the product of coverage fraction and emissivity). Since this approach is based on infrared data only it is not limited to daytime. During daytime, the visible METEOSAT channel provides additional information which may be used for the classification of clouds (Desbois et al., 1982). An earlier multispectral parameterized lookup table approach used four infrared channels from the NOAA-4 VTPR to obtain emissivity, cloud top height, and cloud amount (Bunting and Conover, 1974). As noted previously, the multispectral sounder channel method provides the additional degrees of freedom required to characterize semitransparent clouds (see also Liou et al., 1978; McCleese, 1978).

Houghton and Hunt (1971) suggested a dual wavelength method to discriminate cloud phase (ice/water) and measure ice cloud thickness using passive remote sensing techniques. They also suggested that a near infrared channel (2.6 μm) could provide particle size information. Near infrared channels may be used to distinguish ice cloud from water cloud due to spectral variations in their respective indices of refraction (Blau et al., 1966; Irvine and Pollack, 1968). Current applications include those using TIROS-N AVHRR data (Arking, 1981). Similar techniques applied to near infrared data from SKYLAB, however, give ambiguous results (Curran and Wu, 1982).

4.6.2.3 Summary

The cloud parameter analysis methods discussed above are summarized in Table 4.8 which indicates the required satellite input and supplementary data and the inferred cloud parameters. In general, cloud coverage fraction can be obtained with any of the methods indicated. Cloud coverage fraction and cloud top temperature can be obtained by two channel (vis/IR) operational imagers such as the DMSP OLS. Whenever cloud top temperature is retrieved, cloud top height is available given a temperature profile. However, results from techniques using two channels are degraded by the presence of cirrus cloud. Resolution of the cirrus problem requires the availability of one or more infrared sounder channels either in the water vapor (WV) or CO_2 [IR(CO_2)] regions. The most comprehensive cloud retrievals are potentially obtained from multichannel sounder techniques which are integrated within temperature retrieval procedures. In particular, combined infrared/microwave sounder methods can be used to provide both liquid water content of low cloud and ice cloud thickness. When temperature profiles must also be retrieved, no computational penalty is incurred.

4.6.3 Precipitation

Rainfall (commonly expressed as ground level intensity in mm hr^{-1}) has been inferred from satellite data using visible, infrared, and microwave imagery. An excellent recent review (Atlas and Thiele, 1981) discusses precipitation sampling requirements, available ground truth, satellite techniques, and potential future systems. In this section existing techniques are briefly reviewed.

4.6.3.1 Visible and Infrared Methods

Early approaches to the estimation of rainfall using visible and infrared imagery have been reviewed by Martin and Scherer (1973). Many individual methods are applicable to a wide range of climatic regions. Most techniques are based on readily available satellite imager data. These methods are based on the observation that, since clouds are brighter in the visible and colder in the infrared, they correlate well with areas of precipitation. Further, the intensity of precipitation may be inferred in some cases from the brightness or EBBT. Precipitation itself is not directly measured.

The general approach is to define a statistical relationship, based on an a priori data set, between observed visible or infrared data and the desired precipitation index of the form (Atlas and Thiele, 1981):

$$R = f(I, \Delta I, \sigma_I, A(I), \Delta A/\Delta t \ldots) \qquad (4\text{-}11)$$

where R = rainfall, I = intensity, i.e. brightness for a visible channel and temperature for an infrared channel, A = area, σ denotes standard deviation, t = time and Δ denotes a finite difference. Two classes of techniques can be identified based on the treatment of time dependence. These include: (1) the so-called "indexing" methods which are independent of time and applied to data from instantaneous fields of view (Barrett, 1970, 1973; Follansbee and Oliver, 1975; Kilonsky and Ramage, 1976; Arkin, 1979; Bellon et al., 1980) and (2) "life history" methods which examine time dependent storm development (Scofield and Oliver, 1977; Griffith et al., 1978; Stout et al., 1979; Negri and Adler, 1981). The latter class is generally applied to convective situations only.

Combined visible and infrared methods are attractive for a number of reasons. They might be inexpensively implemented coincident with one of the cloud analysis algorithms described in Section 4.6.2, for example. However, their general applicability over widespread areas has not been adequately investigated.

4.6.3.2 Microwave Radiometry

At microwave wavelengths, precipitation-sized droplets provide a source of atmospheric attenuation analogous to the effect of cloud droplets in the infrared spectrum (Savage, 1978; Falcone et al, 1979). This attenuation mechanism suggests a direct causal relationship between rainfall and microwave atmospheric emission which has been exploited to infer rainfall rate (Weinman and Wilheit, 1981). Furthermore, the presence of precipitation within the field of view of microwave sensors is itself of considerable interest since the quality of resultant retrievals of other quantities is most certainly degraded (Gaut et al., 1974; Phillips et al., 1980; Liou et al., 1981). Microwave sounding brightness temperature must be corrected for rain in much the same way as infrared radiances are for cloud.

The physical basis of microwave remote sensing of precipitation is expressed in Eqn. (4-4) and Table 4.1. To provide a theoretical microwave brightness temperature/rainfall rate relationship, models of both the rain layer (including a rain layer height and thickness) and of rain microphysics are required. Both of these factors, of course, vary synoptically introducing some uncertainty into the retrieval process (see Fowler, et al., 1976). Collectively, these factors determine the atmosphere's scattering and absorption contribution to total satellite incident brightness temperature which is evaluated as a multiple scattering calculation assuming that the rain layer fills the field of view (Savage and Weinman, 1975). The surface background against which the rain is modeled consists of surface emission and surface reflected atmospheric contributions, which must be included. Over the ocean, surface emissivity is low and relatively uniform, providing good contrast for the quantification of precipitation. This approach was applied over the ocean by Wilheit et al. (1977) to data from the NIMBUS 5 Electronically Scanning Microwave Spectrometer (ESMR) operating at 19.35 GHz. Accuracies ±2 mm hr^{-1} and a dynamic range up to about 20 mm hr^{-1} were achieved using ground based radar for validation.

Theoretical calculations suggest that over land higher frequency, dual polarized measurements can distinguish atmospheric scattering due to rain from surface contributions (Savage and Weinman, 1975; Weinman and Guetter, 1977; Huang and Liou, 1983). Studies with the dual polarized 37 GHz ESMR on NIMBUS 6 confirmed these ideas (Rodgers et al., 1979). However, in practice the dynamic range of measurable rainfall rates is so much reduced at 37 GHz that these measurements are virtually useless and considerable ambiguity of surface and atmospheric contributions still exists.

In addition to the problems of specifying appropriate rain models and obtaining measurements over land, another problem arises because the typical horizontal scale of precipitating cloud elements is generally much less than the field of view of the ESMR instrument (~50 km). Integration over a large field of view creates a negative bias which underestimates instantaneous rainfall rates. One suggestion (Wilheit et al., 1982) is to use higher frequencies with correspondingly smaller spatial resolution (e.g., 15 km at 85 GHz). However, the cloud liquid water attentuation at higher frequencies results in an effective lower limit to the detectible rain rate. Multispectral measurements may provide a means to solve this problem.

Within the DMSP, rain rate will be available over ocean and land with simulated rms errors of about 1 and 3 mm hr^{-1}, respectively, from the seven channel SSM/I instrument. Dual polarized measurements will be available at 19.35, 37, and 85 GHz (in addition to a water vapor channel at 22.235). The caveats associated with microwave rain rate inferences described above are generally applicable to this instrument. Additionally, recent calculations (Huang and Liou, 1983) and observations (Wilheit et al., 1982) indicate that the scattering by precipitation-sized ice drastically reduces microwave brightness temperature. This phenomenon was not considered in the development of the SSM/I retrieval algorithm or in the evaluation of its expected rms error performance statistics.

4.7 Surface Properties

The free atmosphere interacts with the Earth's surface radiatively and through boundary layer fluxes of latent heat and momentum (Section 3). A variety of surface properties such as moisture, albedo, temperature, and other indices of surface energy balance may be remotely sensed at window wavelengths. These properties may be useful in providing initial and boundary conditions for NWP forecasts and for verification and tuning of parameterizations used in models. This section summarizes recent research on the retrieval of surface parameters especially in relation to the hydrological cycle.

4.7.1 Soil Moisture

Soil moisture (expressed as percent by weight) may be inferred from microwave measurements (Schmugge, 1978; Burke et al., 1979). Soil moisture is of great agricultural interest and of some military interest (e.g. off-road conditions for heavy vehicles). Soil moisture provides information needed to evaluate surface fluxes of latent and sensible heat and acts as a proxy for antecedent precipitation (Schmugge, 1981; Blanchard et al., 1981). Microwave emission from soils depends on a variety of factors including physical temperature, soil type, surface roughness, and moisture content (Njoku and Kong, 1977; Schmugge, 1978; Wang and Choudhurry, 1981; Tsang and Newton, 1982). For average soil properties and a smooth surface, characteristic emissivities range from greater than 0.9 for a dry soil to less than 0.6 for a wet soil. Roughness generally increases emissivity; local properties such as vegetation also affect soil emissivity (Mo et al., 1982). In spite of the many factors involved, strong correlations have been demonstrated between microwave brightness temperatures at wavelengths of 1.55 and 21 cm and soil moisture in the upper few centimeters of the soil.

Satellite data which might be used to determine soil moisture include that of the DMSP SSM/I sensor. Sensitivity studies indicate that the 19.35 GHz channels will provide surface soil moisture content except over areas heavily vegetated or snow-covered with an rms error of 2.6-3.2%, depending on season, over a dynamic range of 0-25% (Hughes Aircraft Company, 1980).

4.7.2 Surface Albedo

Surface albedo is used directly in the shortwave radiation parameterization in the model. It can also be used to infer snow and ice cover over land and ocean, respectively. A variety of studies have investigated the use of visible multispectral data to infer surface albedos (Salomonson and Marlatt, 1968; Brennan and Bandea, 1970; Raschke et al., 1973; Stowe et al., 1980). Generally these techniques employ some form of the radiative transfer equation to account for atmospheric effects (Odell and Weinmann, 1975; Kaufman, 1979, 1982; Kaufman and Joseph, 1982; Mekler and Joseph, 1983). Subjective methods have generally been applied to discriminate snow covered/high albedo areas using operational visible imagery (McGinnis et al., 1975; Rango, 1975; Matson and Weisnet, 1981; Dewey and Heim, 1982). These operational approaches have generally ignored the physical mechanisms relating snow visible optical properties to hydrological variables such as water content (Bergen et al., 1983). Remote sensing of snow is briefly reviewed by Warren (1982). Microwave measurements at 19.35 and 37 GHz can beused to identify dry snow and snow depth (Schmugge, 1981) as well as sea ice (Gloerson et al., 1974). Retrieval of snow and ice cover will be possible from DMSP SSM/I data (Hughes Aircraft Company, 1980).

4.7.3 Surface Temperature

Window radiances in the infrared (Chahine, 1980; Sidran, 1980) and microwave brightness temperatures (Wilheit, 1980) provide information on ocean and land surface temperature (see McClain, 1979). Infrared methods have focused on the 3.5-4.2 μm window (Shaw, 1970; Smith et al., 1970), the 8-12 μm window (Maul and Sidran, 1973; Prabhakara et al., 1974; Cogan and Willand, 1976) and on multichannel methods (Chahine, 1980, 1981; McMillan, 1980; McClain, 1981; Bernstein, 1982). Error sources for surface temperature inferred from infrared measurements are partial cloudiness in the field of view (McMillan and Dean, 1982), aerosol attenuation (Jacobowitz and Coulson, 1973) and gaseous absorption (Deschamps et al., 1980; Imbault et al., 1981). Multi-wavelength techniques (Bernstein, 1982) and dual viewing angle techniques (McMillan, 1975; Chedin et al., 1982) have been suggested to provide atmospheric corrections. Another approach uses simultaneous data from geosynchronous and polar orbiting satellites (Chedin et al., 1982; Zandlo et al., 1982). Barton (1983) provides a review of surface temperature retrieval methods and a simulated comparison of dual angle and dual channel techniques. He finds that rms errors of 0.1 K are theoretically achievable at 3.7 μm using two viewing angles. Generally satellite infrared surface temperature fields have been found to be biased by 1-4 K (Barnett et al., 1979). However, in a set of carefully analyzed retrievals using NOAA 6 AVHRR data (Bernstein, 1982) accuracies in the 0.5-1.0 K range are reported.

In the microwave region the surface contribution to brightness temperature is due to emissivity as well as physical temperature. Therefore, an emissivity model is required to retrieve surface temperature (Section 4.6.3). Over the oceans surface roughness due to winds, foam, etc. controls emissivity (Wilheit and Chang, 1980; Wentz, 1983). In spite of this problem sea surface temperatures have been retrieved experimentally from microwave radiometers (Hofer et al., 1981; Bernstein, 1982b; Liu, 1983) with accuracies on the order of 1°C. Simulations have been used to investigate the ability of the DMSP SSM/I to retrieve surface temperature. When there is no precipitation, expected rms errors over the ocean are 1.6-2.0 K (Hardy et al., 1981a). Over land, because of variable surface emissivity, errors under operational conditions are expected to range between 3.7 and 4.7 K depending on latitude and season (Hardy et al., 1981b).

4.7.4 Surface Fluxes

Observed fluxes of heat, moisture and momentum could be used to test boundary layer model parameterizations. Although prospects for direct sensing of latent and sensible heat fluxes from satellites are not optimistic (NASA, 1980; Gautier, 1981), related variables such as surface albedo, surface temperature, and soil moisture can be inferred from satellite sensors (see above, and Vukovich, 1983). Measurements of surface winds (Section 4.5) which depend on microwave properties of the ocean surface are actually measurements of surface stress, i.e., of surface momentum flux (Liu and Large, 1981). A number of recent studies have used satellite visible and infrared data with a one-dimensional boundary layer models to infer surface fluxes (Carlson et al., 1981; Price, 1982; Chou and Atlas, 1982). While these case studies have focused on mesoscale phenomena, the application of these methods to large-scale NWP should be investigated. Unfortunately, surface fluxes depend crucially on the stability of the boundary layer. This stability is in turn related to the temperature difference between the ground or ocean and the boundary layer and to the humidity in the boundary layer. As we have seen, these quantities are difficult to measure remotely with high accuracy.

4.8 Implications for NWP

The current operational remote sensing system does not as yet produce data for NWP with the detail and accuracy that is either required or possible. This can be partially remedied by correcting deficiencies in subsystems such as the sensing system itself, the data preprocessing system (e.g., for cloud treatment in temperature retrieval), in retrieval algorithms, and in the assimilation system. We list here some conclusions from this part of our study and some high priority tasks that should be undertaken as soon as possible.

Combined IR and MW instrument systems provide complementary capabilities superior to either component alone. As a byproduct of the combined system concept, cloud properties not available using microwave alone are obtained (see 4.6.2). Studies of combined IR-microwave sounding systems have been performed for AFGL in the past (Weichel, 1976; Burke et al., 1979).

Physical retrieval methods have been demonstrated (see Section 4.4.2.2) to outperform statistical retrievals, especially in cloudy situations. The existence of a serious forward problem discrepancy (thereby precluding the successful implementation of physical retrieval methods) is questionable at this time. Consequently decisions to retrieve temperatures statistically based on this consideration may have been premature. On the other hand, the success of physical retrievals does not imply that there are no forward problem discrepancies (Section 4.3.2). The physical methods currently in use are highly tuned; they are based on physical understanding of the forward problem but they include many empirically determined constants.

Good moisture retrieval algorithms are badly needed, starting with a procedure for obtaining a good first guess. There is usable water vapor information in channels other than those dedicated to moisture determination.

Operational methods can and should be developed for retrieving cloud amount and topography from IR imagery with the use of temperature and window channel fluxes. Attempts should be made to estimate condensed water content and phase, and estimate latent heat release. Cloud parameterization can be tested and tuned through cloud-radiation feedback, both in the model behavior and from the remotely sensed data.

Algorithms should be developed for remotely sensing boundary layer and surface conditions such as surface temperature, ground wetness, snow and ice cover, etc., and for inferring moisture and ground heat sources, surface temperature discontinuity and humidity, and PBL lapse-rate and cloudiness.

Assimilation of remotely sensed asynoptic data is discussed below and development of future sensing systems is beyond the scope of this review. It is important to stress, however, that there is much more useful information in the electromagnetic spectra than is currently being collected. Since lead time for new satellite development is long, it is important that new and better systems of measurements be planned as sensor state-of-the-art improves, and that their potential impact on NWP be estimated by sensitivity tests and impact simulations.

5. CURRENT PROBLEMS IN DATA ASSIMILATION

To begin a numerical weather prediction (NWP) forecast, the complete state of the atmosphere must be specified everywhere in the model domain at a single instant. However, data are observed irregularly in time and space and are incomplete; i.e., not every prognostic variable is observed. Four-dimensional data assimilation methods are designed to solve this problem. Four-dimensional (4-D) assimilation is any analysis scheme which explicitly takes into consideration the time evolution of the atmosphere (see Bengtsson (1975) for a review). For our purposes, and for all practical purposes to date, a 4-D assimilation cycle is a special NWP model run in which, at regular intervals, the model is stopped, the model state is altered to bring it into better agreement with observations, and the model is restarted. The regular intervals referred to may be large (12 h or 6 h), in which case the assimilation scheme is termed intermittant, or small (10 minutes or every time step), in which case the assimilation scheme is termed continuous.

Four-dimensional assimilation goes beyond the notion of time continuity which was used by the earlier investigators: Physical processes may be included in the NWP model; in a continuous assimilation cycle satellite data can be actually assimilated every time step. At any given time it is possible that only a small part of the domain is affected by the data. The effect of the data is then spread and advected by the model until the next alteration time. During this period the mass and motion fields of the model interact so that, for example, even though only temperatures are assimilated, the wind field will be affected as well.

The alteration of the model state has two purposes. The primary purpose is to update the model state so that it agrees better with the current observations. However, simply updating the model state generally causes imbalances which would result in large amplitude gravity-waves in the forecast and rejection of some of the data. The secondary purpose is therefore to initialize or balance the updated model state. In many assimilation cycles updating and balancing are functions which are performed in sequence, but much current work focuses on combining them (see Section 5.3).

In summary, a 4-D assimilation cycle is defined by the updating method, the balancing method and the NWP model. Updating techniques range from simple direct insertion of observed values of the variables at grid points, which was used in several of the early experiments, to complex multivariate three-dimensional optimal interpolation schemes now in use at several modern NWP centers, which combine, in a least-squares approach, observations of several variables within a volume centered on each grid point. Balancing methods range from simple static methods, e.g. forcing the analysis increments to be in geostrophic balance, to much more complex methods, such as nonlinear normal mode initialization (NL NMI) which take into account details of the model formulation and of the particular model state. The NWP models have already been reviewed in general in Section 2 and in more detail with respect to some parameterizations in Section 3.

The plan of the rest of this section is the following: For background we will first describe in some detail two pioneering works by Charney et al. (1969) and by Smagorinsky et al. (1970); these studies are prototypes for the two classes of impact studies and highlight in a simple setting several problems still outstanding in the field of data assimilation. We will then review the current state-of-the-art in updating methods, balancing methods and ways of combining updating and balancing. As we will see, much of current NWP research is concerned with the interactions between the parts of the NWP system; especially those interactions within the assimilation cycle between the forecast model, the analysis method, and the initialization method (Bengtsson et al., 1981; Williamson, 1982; AMS, 1983). Then we will examine the phenomena of data rejection and redundancy while reviewing some observing system experiments. Finally we will discuss implications for our own future work.

5.1 Background

Research in 4-D assimilation was stimulated by the development of satellite observing systems. To take into consideration the asynchronous and incomplete nature of satellite observing systems it was recognized that past observations must be carried along in time by a complete NWP model in a 4-D assimilation scheme. Previously, some operational centers were using the most recent forecast as a first guess in their analysis but not much thought had been given to this procedure and some centers used climatology as a first guess.

In response to the planning needs of GARP, the Global Atmospheric Research Program, a number of observing systems simulation experiments (OSSEs) were conducted. The intention for GARP was to better observe the atmosphere with the ultimate aim of improving our understanding of the atmosphere. To this end a number of novel observing systems were considered, as well as an expansion of the conventional observing network. The OSSEs attempted to determine what observations were needed (wind?, temperatures?, both?), what accuracy was required (1° or 0.5° or 0.25° C rms temperature error?), how many satellites were needed, how should the data be assimilated (continuously or intermittently?, by direct insertion or by optimal interpolation?), etc.

Two types of experiments were conducted - 4-D assimilation experiments and predictability experiments. In these experiments model results are compared to "nature"; "nature" is supposed to be the evolution of the actual atmosphere but is often approximated by a portion of a very long model run. When the experimental model and the "nature" model are the same the experiment is termed an identical twin experiment. Because there is no model error in identical twin assimilation experiments, they obtain reconstructions of the state of the atmosphere which are inherently better than those obtained in analogous experiments using real data or another model as nature. The 4-D assimilation experiments begin with some arbitrary initial state. At regular intervals some data are added to the model and the growth or decay of error is observed during the assimilation cycle. The data are obtained from "nature" and may be contaminated with some observing error, typically assumed to be random. In general the error evolution depends on the initial error, the type of data inserted, the quality of the data, the frequency and method of insertion, details of the model, etc. The purpose of these assimilation experiments was to determine which of these factors are most important and to determine the qualitative nature of the dependence of error evolution on these factors. The predictability experiments begin with an initial state taken from "nature" altered in some specified way. For example, the relative humidity might be set equal to the observed zonal climate mean relative humidity. The evolution of error is monitored during the forecast or ensemble of forecasts. The purpose of these experiments is to determine if some of the variables formally necessary for the initial value problem such as the relative humidity are in fact redundant.

The usual evolution of error in these experiments is depicted in Fig. 5.1. Generally three phases are seen. At first there is a very fast

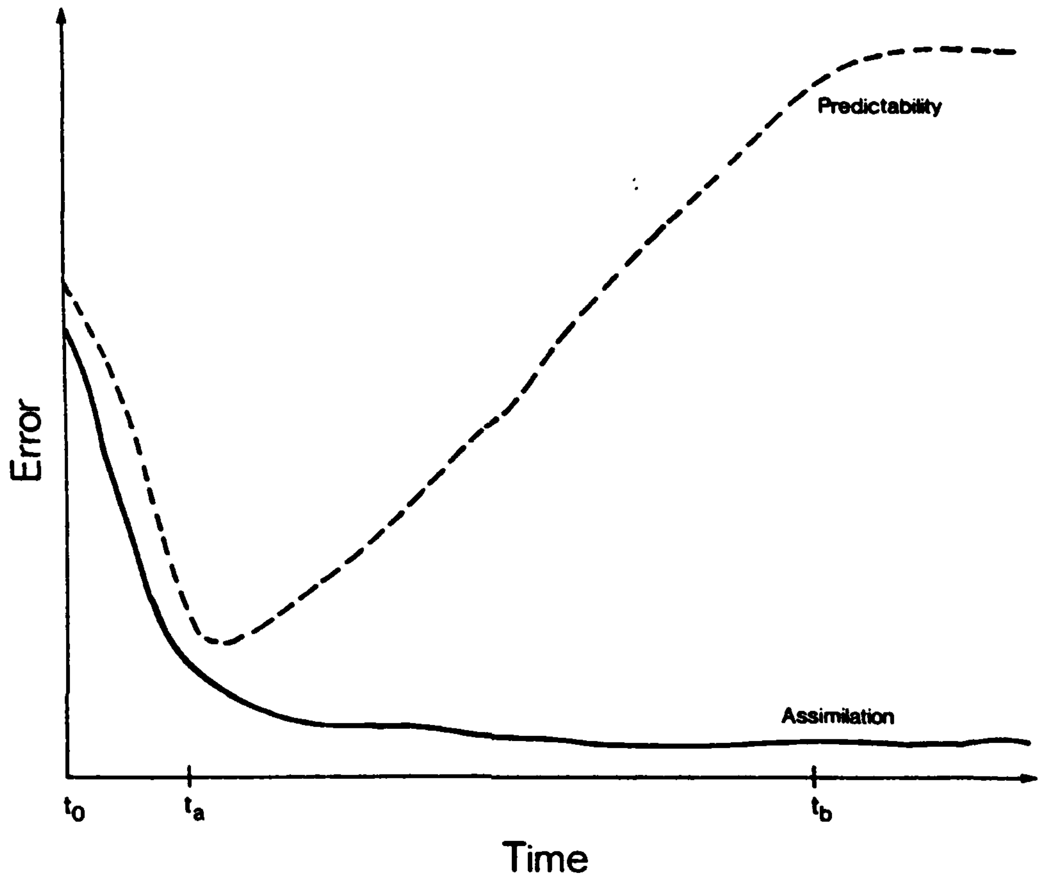


Fig. 5.1 Error evolution. Schematic description of evolution of error during a predictability experiment (----) and during an assimilation experiment (——). There is a very fast adjustment phase ($t_0 < t < t_a$) followed by a relatively slow approach ($t_a < t < t_b$) to asymptotic behavior ($t_b < t$).

adjustment period when the error decays rapidly, as the model tends towards a balanced state. In some cases this is partly due to the dissipation of small scale structures in the error field. After the initial fast adjustment phase there is a relatively slow approach to an asymptotic error level in both types of experiments.

In the assimilation experiments the adjustment is initially rapid but as more and more information is added, each additional piece of information becomes less valuable since the model "remembers" the previously inserted information. Eventually the effect of the initial error is lost and it is only modeling errors and observational errors which prevent a perfect reconstruction. To the extent that these errors have a small constant variance a low asymptotic error level is reached.

In the predictability experiments, if the perturbed variable is in fact redundant, the model will quickly generate a balanced state with a distribution of the variable in question which is nearly correct. For example, the advection of moisture by the correct wind field will shortly produce a reasonable field of relative humidity. By the end of the fast adjustment interval, as the model reaches a state of balance, this effect decreases in importance and error growth becomes apparent.

In assimilation experiments, the model evolution is discontinuous at the insertion times, whether they are every 10 minutes or every 6 hours. We may think of the insertion of data as causing a "shock" to the system or model. One possible system response is to damp out the shock and return to its original unperturbed state. This phenomenon is termed data rejection. Typically the insertion of data produces imbalances in the model which are usually evidenced by the generation of large amplitude gravity waves.

Charney, Halem and Jastrow (1969) were the first to clearly formulate the 4-D assimilation problem: One would like to define a complete 3-D state of the atmosphere to be used as initial data for a numerical forecast. Since the requisite observing system is not feasible, they asked whether observations distributed in time and space can be used to estimate the current state of the atmosphere. Of course continuity in time was always an ingredient in objective analysis, but in this paper the special asynchronous and incomplete nature of satellite observations is addressed.

Possible GARP data systems were considered, raising the questions of whether the wind field and surface pressure may be constructed from temperature observations alone and whether the prognostic variables below the cloud tops may be inferred from observations of the variables above the cloud tops and observations of the surface variables. Because of the dynamic coupling between variables of different types and at different locations, it was expected that these questions would be answered in the affirmative. This completeness issue is discussed at some length by Ghil (1980), Bube and Ghil (1981) and Talagrand (1981a).

In support of these conjectures, Charney et al. (1969) describe and present results from some identical-twin experiments with the dry Mintz-Arakawa two-layer global pressure coordinate primitive equation GCM using 7° x 9° latitude-longitude resolution. In the first series of experiments temperatures with or without random observing errors were inserted globally at regular intervals, Δt, i.e., intermittently. The best reconstruction of winds was obtained for $\Delta t = 12$ h; for shorter Δt two things happen: First, more information is added to the temperature field; by itself this would improve the reconstruction. Second, there is less time for the model winds to adjust to the temperature field between the shocks of data insertion, which are evidenced by large amplitude gravity waves. The reduction of rms wind error was relatively smaller at the equator than elsewhere. Presumably away from the equator the wind and mass field are strongly coupled quasi-geostrophically while near the equator they are not. However, in absolute terms, the asymptotic error levels at 49° and 0° are similar. In the second series of experiments the wind and temperature in the lower layer were reconstructed by inserting surface pressure and upper layer wind and temperature. For no observational error and global insertion at $\Delta t = 12$ h the accuracy of the reconstruction is impressive. Further experiments of this type were carried out by Jastrow and Halem (1970) and Kasahara and Williamson (1972).

Smagorinsky, Miyakoda and Strickler (1970) performed predictability experiments with the 9-level moist hemispheric N40 σ-coordinate primitive equation GFDL model to determine whether observations of surface pressure, relative humidity, boundary layer temperature or boundary layer velocity are in some sense redundant. The variables tested by Smagorinsky et al. were chosen because these "variables are relatively difficult to measure or to determine with high accuracy either for technical reasons or because of the variability of the quantity." It was therefore hoped that they would not be too important in the prediction problem. The smaller-scale variability of the boundary-layer and humidity variables suggests, however, that a finer resolution model might alter the conclusions of these experiments. For each of these variables, a specific perturbation was applied to the initial conditions of 1200 GMT 9 January 1964 and a 14-day integration was performed and compared to the control integration started from the unperturbed initial conditions. To test the effect of surface pressure a constant value of z_{1000} was used for initial conditions; to test the effects of relative humidity and boundary layer temperature the observed values were replaced by their zonally averaged values; to test the effect of boundary layer velocity the observed values were replaced by zeros. In the context of these experiments the boundary layer is taken to be the lowest two σ-levels. The evidence presented shows that these variables rapidly adjust towards their values in the control experiment and are redundant, at least for the model which was used. Even after 96 h the topography of z_{500} is very similar in all five experiments. In all the tests the major differences disappear within 24 h and after this period the first simulated week for each forecast is similar to the control. Beyond one week differences grow and the solutions no longer resemble each other. Of course the specific results of these experiments are model dependent. However, it is evident from the relative operational success of the equivalent barotropic model that 500 mb heights alone determine to a large extent the future evolution of the 500 mb surface and further that the patterns of 500 mb heights are well correlated with weather phenomena at other levels.

Therefore the general results of these experiments appeared very plausible. However, more recent experiments with the GFDL model using conventional NMC initial conditions (Miyakoda et al., 1978; Sirutis et al., 1980) show that it takes 2.5 to 3 days of model integration for the tropical rainfall to appear ("spin up"). This experiment included convective precipitation; the Smagorinsky et al. (1970) experiment did not (Lejenäs, 1979). Thus, the initial specification of moisture and divergence is very important for the tropical precipitation forecast. In the extratropics, a spin up time of 15 h is reported by Lejenäs (1979); his experiments show convincingly that both vertical velocity and moisture must be specified consistently to obtain reasonable initial precipitation rates. Unfortunately, Lejenäs did not examine the effect of his moisture initialization, which we discuss further in Section

5.3, on the forecast of quantities other than the hemispherically averaged precipitation rate.

5.2 Analysis Methods

In the identical twin experiments of Charney et al. (1969) "nature" was conveniently "observed" at the model's grid points. This eliminated the need for an analysis, but in general, during an assimilation cycle the model's representation of the atmosphere is updated by performing an analysis. Analysis methods determine either a spectral or gridded representation of atmospheric variables from observations distributed irregularly in space. The observations may, in general, include quantities different from the one being analyzed; in this case the scheme is multivariate, otherwise it is considered to be univariate. For example, a multivariate scheme may use winds, radiances and temperatures to analyze temperature. Usually the analysis depends linearly on the data. Often this dependence is local, i.e., the analyzed value at a grid point depends only on neighboring observations. The analysis methods fall into three classes which we will call the successive correction method (SCM), optimal interpolation (OI) and the variational analysis method (VAM).

The successive correction or Barnes-type analysis (Bergthorsson and Döös, 1955; Cressman, 1959; Barnes, 1964; Lipton and Hillger, 1982; Baker, 1983) forms a succession of weighted averages of a background field and the observations. This method is univariate and usually two-dimensional. During each scan of the field the previous analysis is used as the background. After each scan the neighborhood which contributes observations to each grid point shrinks. The background for the first scan may be a forecast (as in an assimilation cycle) or climatology. This method preserves detail (i.e., small scales) in data rich areas and produces smooth fields in data poor areas. It is very efficient. The SCM may be considered an approximation to OI (Seaman, 1983) and comparison to OI may be used to estimate to what extent the SCM is suboptimal and how to tune the SCM.

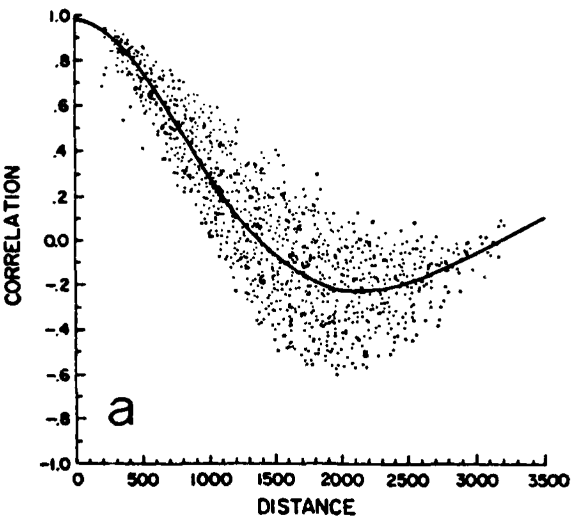
The OI or Gandin-type analysis is in essence linear regression of the grid point values in terms of the observations (Eliassen, 1954; Gandin, 1963, 1969; Rutherford, 1972; Schlatter, 1975; Lorenc, 1981). This allows considerable flexibility since OI can be multivariate and three-dimensional (Williamson et al., 1981). Since it can be multivariate, OI can, in principle, directly assimilate radiances. Betout and Yakima (1982) have experimented with this approach. They calculated the necessary covariances by linearizing the equation of radiative transfer and used the operational clear column radiances. Forecast skill was not improved. A nonlinear VAM for the direct assimilation of raw radiances and cloud filtering has been proposed by Hoffman (1983).

In the VAM (Sasaki, 1970; Wahba and Wendelberger, 1980) the grid point or spectral coefficient values are determined which best fit the data. In the original Sasaki approach an SCM was used to first grid the data. There is a very wide range of numerical methods available to solve this problem (above references and Hoffman, 1982; Halberstam and Tung, 1983; Halem and Kalnay, 1983). Every OI may be stated as a variational problem (Kimeldorf and Wahba, 1970; Wahba, 1982). The VAM is very flexible in allowing auxiliary constraints to be included (see below Section 5.3); indeed this was the original motivation for its development by Sasaki (1958), Lewis (1972) and their co-workers.

All of these methods require in some way estimates of some of the statistics of the quantities observed and analyzed. Typically one must specify relative error magnitudes of the observation and spatial correlations of the true field. For example, in operational OI schemes observed spatial correlations of height fields are fit by a Gaussian shape and the result differentiated to give geostrophic height-wind and wind-wind correlations (Schlatter, 1975; Julian and Thiebaux, 1975). Typical correlations for 500 mb heights and temperatures are shown in Fig. 5.2. Recent work (Balgovind et al., 1983) suggests using a best fit Bessel function.

We now consider in some detail (1) humidity analysis schemes, (2) the spatial variability of moisture, (3) the use of some non-conventional moisture variables and (4) the sensitivity of short range forecasts to improved moisture analysis. Currently, operational moisture analyses do not account for some of the special characteristics of atmospheric water vapor and do not use some potentially useful, but non-conventional observations.

The objective analysis of the moisture field requires special care because relative observational errors are larger, length scales are smaller and the variation of specific humidity, which is the model variable, is very substantial both from surface to tropopause and from pole to equator. Van Maanen (1981) presented some statistics for water vapor mixing ratio, r, which



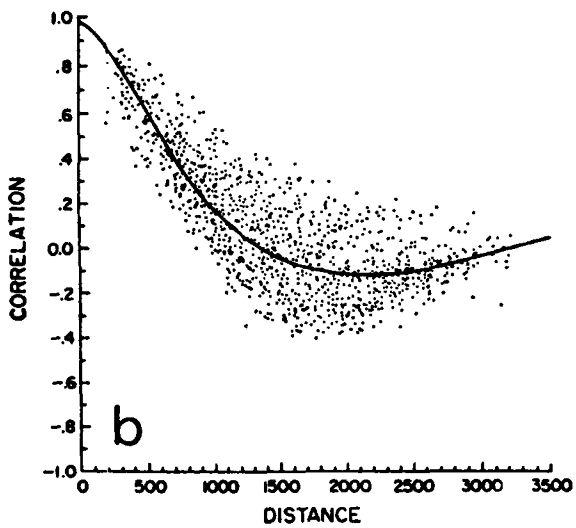


Fig. 5.2 Height and temperature correlations for Topeka, Kansas. Correlations are plotted as a function of separation distance (km) for heights (a) and temperatures (b) at 500 mb. Curves plotted are the best fit damped cosine (from Julian and Thiebaux, 1975).

we reproduce here. For winter and summer 1977 and 1978, he calculated the correlations between the observations of the radiosonde station at de Bilt and of other European radiosondes (Fig. 5.3a and Fig. 5.3b). The results were fit to the form $\rho_0 \exp(-bs)$ where s is separation distance (Table 5.1). Note how quickly the variance falls off with height. The product of ρ_0 and the observed variance is the "true" variance of the field at the scale of the observing network. The remainder of the variance is due to errors and variance on scales smaller than the observing network. Van Maanen studied the correlations of surface observations of mixing ratio in order to determine the cause of the very low value of ρ_0 for 850 mb in summer (Fig. 5.3c and Fig. 5.3d). The results show that there is considerable variability at scales small compared to typical radiosonde separation but large enough (>10 km) to be resolved by the surface network.

By comparing Figs. 5.2 and 5.3 we can see how the variability of the different quantities is distributed with respect to scale. In the winter very little small scale variability, i.e., variability at scales below the resolution of the radiosonde network, is evident. Height is characterized by larger scales than temperature which in turn is characterized by larger scales than mixing ratio. For example, at 500 km separation, the autocorrelations for these quantities are approximately 0.8 for height, 0.65 for temperature and 0.5 for mixing ratio. When comparing these autocorrelations, note that there are differences in place, year, vertical level and definition of the mean value used in the calculations. Unfortunately not many figures of this sort have been published. For small scales (20-200 km) Barnes and Lilly (1975) have presented consistent calculations of correlations from a single data set. Some of their results are shown in Fig. 5.4, which clearly illustrates that humidity varies relatively more at smaller scales than temperature. Norquist and Johnson (1982) also investigated the statistics of moisture - in particular, specific humidity. Their work is difficult to interpret since in areas of poor data coverage they studied the NMC operational analyses which have significant shortcomings, and in areas of good data coverage, scales not resolved by the radiosonde network were absent.

There does not seem to be a clear cut choice as to which moisture variable should be analyzed - specific humidity, mass mixing ratio or relative humidity. On the one hand it is desirable to analyze relative humidity, which

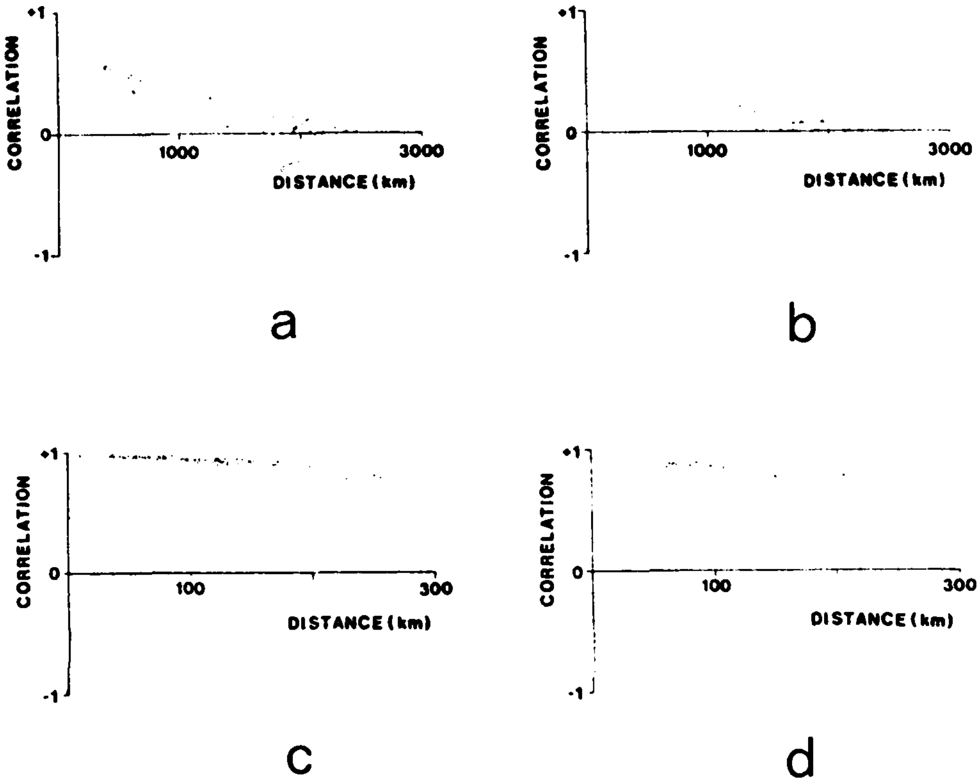


Fig. 5.3 Mixing ratio correlations for the Netherlands. Results are plotted versus separation distance for radiosonde observations at 850 mbar for de Bilt, for the winter (a) and summer (b); and for surface observations for the Netherlands for the winter (c) and summer (d) (from van Maanen, 1981).

Table 5.1 Water vapor mass mixing ratio statistics. Values of ρ_o and b obtained by fitting the function $\rho_o \exp(- bs)$ to the correlations. The observation variance is the sum of the variance of the true humidity and the variance of the error (from van Maanen, 1981).

				Variance ($\times 10^6$) of		
Season	Level	ρ_o	1/b(km)	Observation	True humidity	Error
Winter	850	0.95	730	1.81	1.73	0.085
	700	0.74	550	0.85	0.63	0.22
	500	0.93	480	0.11	0.10	0.008
Summer	850	0.54	990	2.95	1.60	1.35
	700	0.86	440	1.87	1.61	0.26
	500	0.78	330	0.24	0.19	0.054

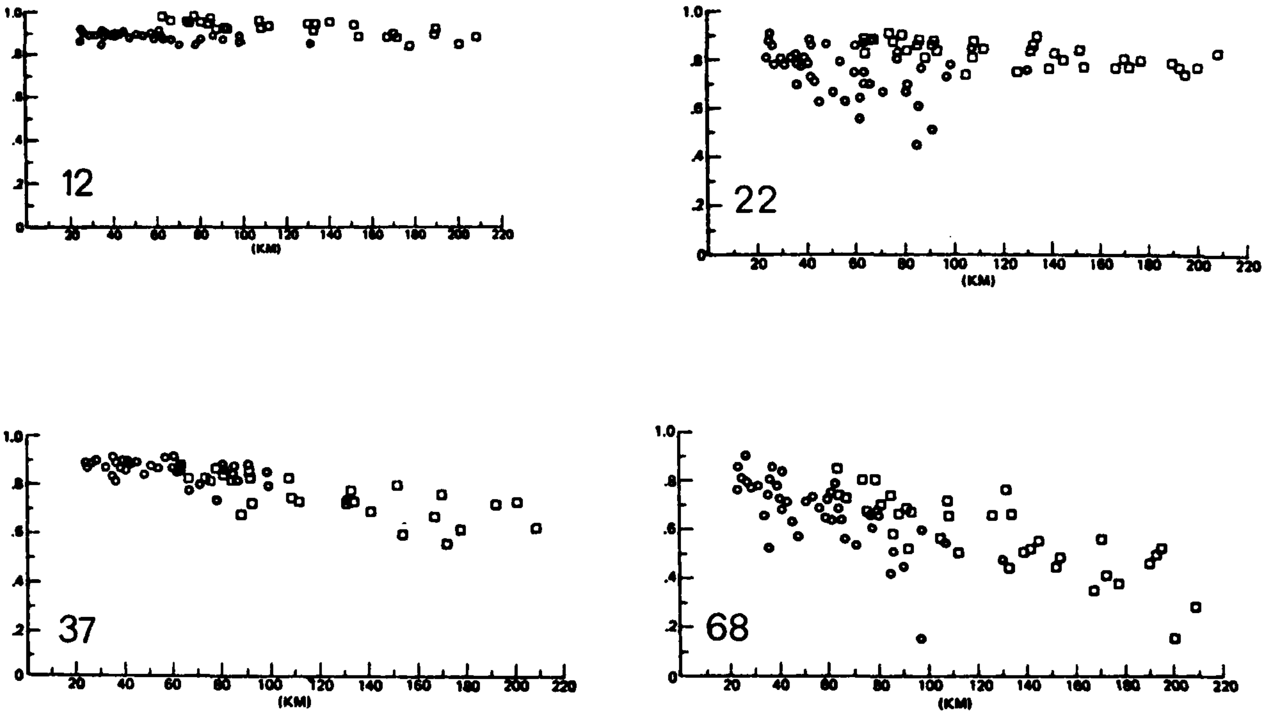


Fig. 5.4 Temperature and mixing ratio correlations for Oklahoma. Correlations as a function of separation distance are shown for temperature (top) and mixing ratio (bottom) under nonstormy (left) and stormy (right) conditions. Nonstormy and stormy conditions were identified by the absence and presence of convective radar echos respectively. Data are for spring seasons, 1966-68 at 1500 m above ground level. The numbers in the lower left corner of each plot is an estimate of the percent of total variation at scales less than 200 km. Diamonds represent 1966-67 data; open circles, 1968 data (from Barnes and Lilly, 1975).

does not vary by orders of magnitude and which is most directly related to cloudiness. Also since all observations and forecasts of relative humidity will be in the range 0 to 1, any analysis scheme which is of the form of a weighted average will yield realizable moisture values. But relative humidity is not the model variable and depends on both specific humidity and temperature. To our knowledge all objective analyses of moisture have been univariate. The ECMWF (Lorenc and Tibaldi, 1980; Bengtsson et al., 1982a) and GLAS (Baker, 1983) use SCM analyses of layer mean water content and relative humidity respectively in their FGGE four-dimensional analyses. NMC (McPherson et al., 1979) operationally uses an OI analysis of specific humidity. GFDL analyzes dewpoint temperatures (Sirutis et al., 1980). Special consideration was given to the problem of moisture analysis by Atkins (1974) in a SCM analysis of relative humidity and by van Maanen (1981) in an OI analysis of mixing ratio. Atkins' analysis preserves frontal structure in the first guess field by using first guess gradient information to give to observations larger and smaller weights, respectively depending on whether they fall parallel or perpendicular to first guess isopleths.

Some observations of moisture from satellites are not commonly used because they are observations of some quantity which depends on humidity but which cannot be inverted to yield humidity profiles. This type of observation includes column integrated water vapor, column integrated liquid water, precipitation rates, and cloud imagery (see Section 4.6 for details). Here we discuss an interesting method which uses column integrated water vapor, another which uses imagery and ground-based observations of the occurrence of precipitation, and an operational system, the Air Force three-dimensional nephanalysis (3DNEPH) which may be considered a humidity analysis as well as a cloud analysis; the use of precipitation rates is discussed in Section 5.4.

Haydu and Krishnamurti (1981) combined microwave spectrometer measurements of column integrated water vapor, radiosonde profiles of specific humidity, and conventional surface observations of specific humidity to obtain high resolution, three-dimensional specific humidity analyses. They used the vertically integrated radiosonde values to calibrate the microwave inferred measurements while preserving the gradients of the microwave measurements. They then assumed that the vertical distribution of specific humidity decays exponentially from the surface value, given by the analysis of conventional surface observations, to a value of zero above 500 mb. This assumed profile and the constraint of the vertical integrals is enough to specify the humidity everywhere.

Wolcott and Warner (1981) combined satellite imagery and present weather observations to identify grid boxes which should be saturated using the following method: They first matched a separate temperature analysis to the satellite observed cloud top temperatures in order to assign cloud top pressures. Clouds were assumed to exist at one model level unless the surface observer's report of present weather indicated precipitation, in which case the cloud was assumed to extend to the surface. All cloudy grid boxes identified in this way were assumed to be saturated. The other grid points were analyzed from radiosonde and surface data either by an ordinary SCM or by Atkin's method. Compared to a control analysis which used only radiosonde data the augmented analysis has much sharper features and produced much better precipitation rates during the first 4 h of the forecast. In these experiments the Tarbell et al. (1981) initialization procedure was used.

The AFGWC cloud forecast system is a unique operational system (Fye, 1978; see Mitchell and Warburton (1983) for a summary); part of this system is the 3DNEPH which uses satellite observations of cloud as a principal data source (Section 4.6.2). The cloud forecast model uses trajectories from a separate dynamic model to advect water substance, assuming adiabatic conditions. The model conserves water except when the relative humidity reaches 100%, in which case it rains; rain is evaporated as it falls through unsaturated layers. The initial specification of moisture is made from the cloud analysis using a relationship like (3-1) which is described graphically by Fig. 3.3. The prediction of moisture is also translated into a cloud forecast by this relationship. The cloud analysis is described in detail by Fye (1978); some of its attributes are: the first guess is persistence; the analysis cycle is 3 h; the horizontal resolution is 47.6 km; there are 15 vertical layers; both total cloud and layer cloud are analyzed. Cloud height is assigned by comparing the IR EBBT to a separate temperature analysis. The coverage obtained from the satellite imagery has such high resolution that data is simply combined in boxes; no interpolation is needed. Conventional data is spread to neighboring gridpoints using an SCM-type weighting function.

It has generally been perceived that the moisture analysis was not very important and that for forecasts of 24 h and longer only very substantial improvements in the moisture analysis would have any significant effect on the forecast. However, in several studies using regional models it has been shown that the short term precipitation forecast is sensitive to the initial moisture field. Similar results have recently been reported for global models at GLAS (Halem, private communication) and at GFDL (Miyakoda, private communication). Also the impact of latent heating experiments discussed in Section 3.6 help to delineate the sensitivity of the model forecasts to moisture.

Atkins (1974), using the UKMO fine mesh PE model, found that her improved SCM analysis of relative humidity had very little effect on the 24 h forecast of surface pressure, but yielded improved precipitation forecasts. Perkey (1976) found that by enhancing the initial moisture field by making it consistent with satellite-observed cloudiness he obtained better precipitation forecasts from his limited area PE model. These experiments and several additional experiments reported by Perkey (1980) show that the short term precipitation forecast is sensitive to small scale features in the initial humidity field. Better vertical resolution in the PBL also improved Perkey's (1976) results. Perkey's (1976) model carries cloud water and rain water as prognostic variables, but he had no initial data for these variables. He initialized these quantities as zero and found that his model produced no rain for several hours. Mills and Hayden (1983), using the ANMRC regional PE model with a horizontal resolution of 67 km, found significant improvement in forecasts of humidity and vertical motion when high resolution satellite observations of temperature and humidity were used to improve the analysis in comparison to a control experiment based on the LFM analysis.

Mitchell and Warburton (1983) conducted sensitivity tests of the humidity forecasts of the NMC spectral model (Sela, 1980) to the initial specification of moisture. They used a total of nine cases. In the first set of forecasts the conventional NMC Hough analysis of moisture was used; in the second set, the Air Force cloud analysis was used for initial data by translating it into a humidity analysis as described above. When translated into and verified against total cloud cover, the second set of forecasts was substantially better for the first twelve hours, and somewhat better out to 48 hours.

We conclude by listing a number of outstanding and troubling issues common to all current objective analysis schemes.

1) Objective analysis techniques essentially try to find an "optimal" solution. But is there a best way of defining "optimal"?

2) Objective analysis techniques smooth the data. What can we do about fronts?

3) Our a priori knowledge of the atmosphere goes beyond means and covariances. For example, when a satellite observes half of a storm, can we objectively make use of our knowledge of "typical" mid-latitude cyclone structure, to infer the structure of the other half.

4) Retrieval of the meteorologically relevant information from remotely sensed electromagnetic parameters depends on the a priori information used. (By "information" we mean statistics, other data, models, etc. Since models are continuously being improved the retrieved meteorological information can never be considered final.) How can all possible information be combined into the retrieval process?

5) Objective methods tend to blend data from different sources. When two data sources conflict a human will "draw" for the more reliable one. Often an average of several reasonable analyses will not be a reasonable analysis but rather an approximation to climatology. Can this problem be resolved within an objective framework?

6) Data with large errors, but which pass the gross error checks, can strongly affect the analysis. There are some reasonable approaches when the data is dense (Fleming and Hill, 1982) but are there good objective methods to detect isolated outliers?

7) Gross data errors due to coding or communications errors can sometimes be corrected by a human. Can an "expert" artificial intelligence computer program do the same?

5.3 Balancing Methods

It has been found that the analysis should be balanced. Balancing the initial state is desirable for several reasons. First, balancing eliminates unrealistic high amplitude gravity waves which might have a deleterious effect on the future forecast. To the extent that the gravity waves are a linear phenomena superimposed on the meteorologic motions they are only a computational nuisance; however, it is thought that the condensation forced by the vertical motion field of the gravity waves may affect the forecast beyond 12 h. There is no evidence supporting this conjecture, although it is physically reasonable. Note, however, that the GLAS model, which has forecast skill comparable to other current models, uses no initialization other than a Euler backwards time scheme (Halem et al., 1982). Second, when not all variables are observed the balance condition may be used to infer the unobserved variables. Third, the atmosphere is balanced, so imbalances in the analysis reflect data inaccuracies. Thus proper balancing can be used to filter data errors. This approach does require that the balance condition used be quite accurate. As discussed above, balancing is particularly important in an assimilation cycle because data insertion shocks the model.

We know of no operational approach in which the moisture fields are balanced with respect to the other fields; even though it is natural to think that relative humidity and vertical velocity should be well correlated. In fact, Lejenäs (1979) found there was negligible correlation between relative humidity and vertical velocity in his PE model, unless he restricted the sample to points with time-averaged vertical velocities of large magnitude. This observation led Lejenäs to the following somewhat ad hoc, but successful moisture initialization: the relative humidity of each parcel was adjusted as if it were displaced vertically by an amount proportional to the time-averaged vertical velocity. In this case the initial hemispherically-averaged precipitation was very realistic if the vertical velocity field was properly specified. While Lejenäs took the approach of adjusting the humidity field to the vertical velocity field, another possibility is to adjust the vertical velocity field (or equivalently the divergent windfield) to the humidity field. This latter approach was taken by Tarbell et al. (1981). Tarbell et al. used observed areas of precipitation to force a diagnostic calculation of the divergent wind. This initialization procedure supported the continuance of the observed precipitation at the start of the forecast and produced better precipitation forecasts for the first 3 h of the forecast period compared to a control forecast.

Balancing techniques range from requiring a geostrophic balance between the analysis increments of height and velocity (Kistler and McPherson, 1975) to nonlinear normal mode initialization (NL NMI) (Daley, 1981). All these methods may be considered filters which suppress high frequency gravity waves. They all depend either explicitly or implicitly on a clear separation of time scales in the atmospheric motions between meteorological "signal" and gravity wave "noise." This idealized separation is not obtained in reality; there is no clear frequency separation for some of the Rossby and gravity waves which have large amplitude in the tropics. This causes substantial problems especially with respect to the Hadley cell (i.e., the meridional overturning in the tropics). It turns out that some gravity wave modes of the atmosphere have long periods and are needed to represent the Hadley cell. It may be that these gravity wave modes are forced by and in turn act to organize the latent heat release in the Intertropical Convergence Zone.

All these filtering methods remove degrees of freedom from the representation of the atmosphere. If geostrophy is specified then winds and heights can not be independently specified; for example, balanced winds may be found from the heights. If an NMI is used then one only needs to specify the Rossby mode amplitudes to deduce the correct gravity mode amplitudes. Direct application of these methods are examples of simple balancing; in simple balancing some information is not used. Variational techniques on the other hand are commonly used to find a balanced state which is "closest" (in some specified sense) to all the data (Sasaki, 1958, 1970; Sasaki and Goerss, 1982; Daley, 1978; Halberstam and Tung, 1983). The variational approach is very attractive in principle, but in practice there are substantial technical difficulties in applying it. Also, the balance conditions are often not precise or correct enough. For example, since the balance conditions used are all approximations, a perfectly observed atmospheric state will in general be altered by balancing.

Before the work of Machenhauer (1977) and Baer and Tribbia (1977) who showed how the normal modes of a PE model could be used to balance the initial conditions, a number of efforts were made to develop diagnostic balance relationships from scaling arguments (Houghton et al., 1971 and references therein). These early methods were not successful in suppressing gravity waves in the forecast. These methods may now be viewed as approximate NMI methods and we now know that the balances required to eliminate large amplitude gravity waves is excruciatingly exact - so exact in fact that details of the model finite differencing must be accounted for. In the remainder of this section we will concentrate on describing recent simple and variational NMI studies. They are sometimes referred to as unconstrained and constrained NMI since simple NMI is not constrained by data, but variational NMI is constrained by data.

5.3.1 Simple Normal Mode Initialization Techniques

Normal modes may be calculated for a continuous atmosphere (Kasahara, 1976) or for a particular model (Williamson and Dickerson, 1976; Ballish, 1980 and references therein). The normal modes are found by linearizing the governing equations about a uniform basic state of no motion. With this simple basic state the problem separates into vertical and horizontal eigenvalue problems. The vertical eigenvalue gives the equivalent depth and the horizontal eigenvalue the frequency. A normal mode is a vertical structure function multiplied by a sine or cosine function of longitude multiplied by one of a set of meridional structure functions. For each vertical structure function and each longitudinal wavenumber there is a set of meridional structure functions.

Because each numerical model has its own discretization error, the modes calculated for a model differ from modes calculated for a continuous atmosphere. The latter we will call Hough modes. The difference between the model and Hough modes is substantial for modes with small scales. Finite differencing also creates model modes which have no corresponding Hough mode. Such modes are termed computational (Baer and Sheu, 1983). As a consequence, it is necessary to use the exact model modes to initialize a particular model properly. Note, however, that when used in conjunction with dynamical methods (e.g., Nitta and Hovermale, 1969) knowledge of the Hough modes may be sufficient (Bratseth, 1982).

The simplest use of the normal modes is to eliminate the unwanted gravity wave modes from an atmospheric state. This technique is called linear NMI or gravity wave zeroing. In brief, for a discrete model, the entire set of normal modes form a complete basis and any model state may be expressed as a linear combination of the modes. When the modes are orthogonal, as is usually the case, coefficients in this representation are found by calculating an inner product of each normal mode and the given state. This process is known as projection. The gravity wave coefficients are then effectively zeroed and the linearly balanced state is obtained by forming the sum of all the retained normal modes each multiplied by its coefficient.

While gravity wave zeroing eliminates all gravity waves from the initial state, nonlinearities in the complete model produce large time tendencies for the gravity wave modes from such initial states. A solution to this problem (Machenhauer, 1977) is to zero the gravity wave tendencies instead of the gravity waves themselves at the initial time. This results in a nonlinear equation which is solved by iteration, taking the nonlinearities as known at each iteration. The meteorological modes or Rossby modes are not altered by this method. Convergence is not assured. A more general method was obtained by Baer and Tribbia (1977) by using a two time scale analysis, assuming that nonlinearities are weak. The bounded derivative method (Kasahara, 1982) is conceptually equivalent to the Baer-Tribbia scheme. Theoretically the Baer-Tribbia scheme can be used to eliminate all fast time behavior; in practice it is difficult to implement. This scheme also does not alter the Rossby modes. These methods are called NL NMI methods. NL NMI has been successful in suppressing high frequencies and providing compatible initial vertical motion fields (Williamson and Temperton, 1981; Daley, 1979). However, simple NL NMI produces adjustments which are sometimes larger than the uncertainties in the analysis (Williamson and Daley, 1982).

5.3.2. Variational Normal Mode Initialization Techniques

In the simple linear and nonlinear NMI schemes the observed Rossby mode amplitudes are unchanged. The resulting model state is in balance but may not agree well with the observations. An alternative variational approach is to find Rossby mode amplitudes which yield a balanced model state and which fit the data well. For example, the Flattery (1971) analysis method which was used operationally at NMC is the simplest scheme of this type. The basis of the variational approach is that each possible choice of Rossby mode amplitude has a (linearly or nonlinearly) balancing set of gravity mode amplitudes. Together all these modes define a model state, which we compare to the observed data. We want to find the set of Rossby modes which gives the balanced model state which best agrees with the data (Daley, 1978; Tribbia, 1982; Daley and Puri, 1980). Within the slow manifold conceptual framework of Leith (1980) a simple linear NMI projects an observed state onto the Rossby manifold holding the Rossby modes fixed; a simple NL NMI does the same onto the slow manifold; a variational NL NMI finds the state on the slow manifold closest to the observed state (Fig. 5.5).

Variational NL NMI is a very appealing solution to the initialization problem because shocks to the model are avoided and the adjusted state agrees

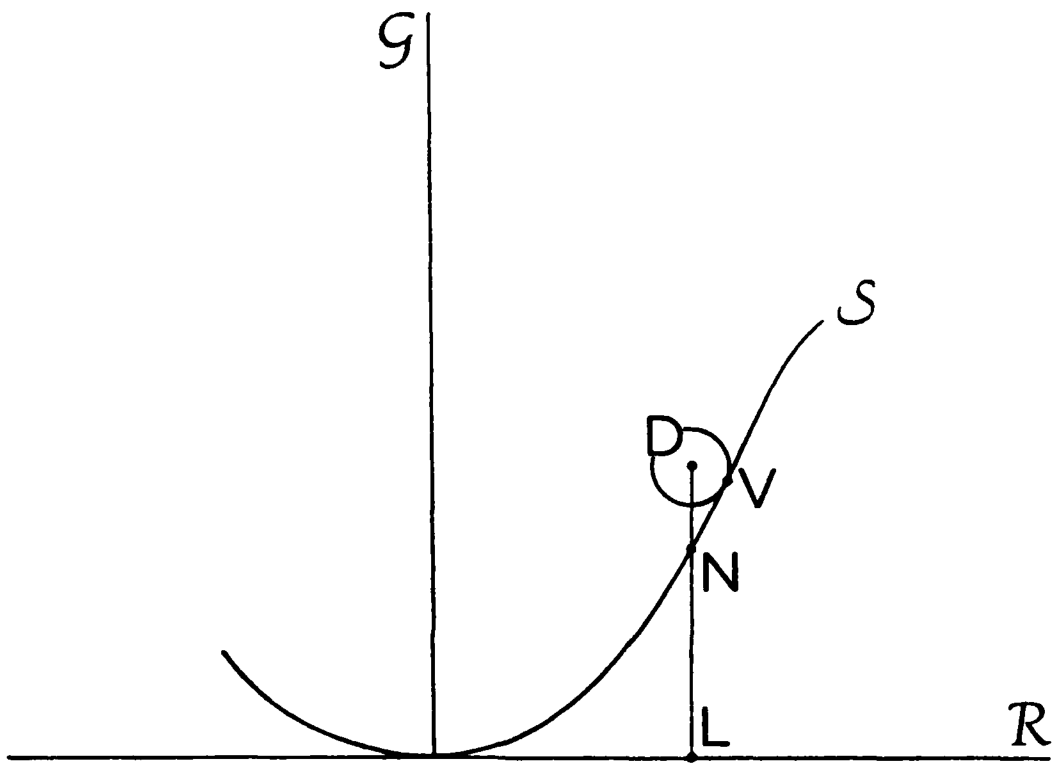


Fig. 5.5 Geometric interpretation of NMI. The manifolds indicated are Rossby (R), gravity (G), and slow (S). An arbitrary initial state defined by data (D) is projected by linear and NL NMI onto the points L and N respectively. Point V is obtained by variational NL NMI (see Leith, 1980 for details).

with the data. However, implementation of this method in an operational system is a difficult problem. Another difficulty with the variational approach is the choice of weights (Wahba, 1982; Puri, 1983b). A solution may be to iterate a standard analysis and simple initialization using the current initialized state as the first guess for the next analysis (Williamson and Daley, 1982; Barker and Rosmond, 1983). The expected error covariances for the first guess should change with each iteration.

5.3.3 Problems with Normal Mode Initialization

There are several outstanding problems with the use of NMI (Daley, 1981). First, in the tropics there is no good time scale separation between meteorologically significant and gravity wave modes, so it is difficult to decide which modes to initialize. A resulting problem, for example, can be seen in the ECMWF FGGE analyses, which use the Machenhauer scheme, in which a very weak Hadley circulation results. Recent work on avoiding this problem has suggested initializing only modes with periods less than 6 h (Puri, 1983a; Stern and Ploshay, 1983) and initializing only the analysis increments (Puri et al., 1982). These suggestions, which tend to leave the gravity wave modes unaltered during the assimilation cycle updating, are reminiscent of an experiment by Dey and McPherson (1977) in which the divergence field was not directly altered by data insertion during an assimilation cycle. Dey and McPherson found that this procedure produced only a neutral impact. Second, there may be nonelliptic regions where convergence of the Machenhauer scheme is not possible (Tribbia, 1981). For both these problems it may be necessary to include the effects of latent heat release in the tropics in the NMI. Third, for gravity modes associated with small equivalent depth the Machenhauer scheme does not converge so modes with equivalent depths smaller than a critical equivalent depth may not be initialized. The Baer-Tribbia scheme has not been operationally implemented because some of the required calculations are so complex. Fourth, initial rainfall rates are not much improved by NMI although vertical velocities are; this is probably due to a poor initial moisture analysis. Fifth, the initialization procedures do not work well in regions of high topography (Ballish, 1983). Recently, Tribbia (1983) has extended the Machenhauer scheme to higher order by using the full model to calculate the Taylor expansion of the nonlinear terms. This new approach is equivalent to the Baer-Tribbia scheme but avoids the computational difficulties of the Baer-Tribbia scheme. Since the full model is used, latent heat effects are automatically included.

5.4 Dynamic Methods

Since the atmosphere is balanced it is no surprise that an undisturbed model will approach a balance with very little gravity wave energy present. That is, both the atmosphere and a reasonably constructed model dissipate gravity waves. As a consequence a variety of methods have been developed to use the dissipation of the model itself to obtain a balanced state while inserting data. Generally the data is first analyzed to grid points using one of the schemes presented in Section 5.2. In some cases the gridded data are directly inserted, usually more than once; in other schemes a forcing function proportional to the difference between model state and gridded data are appended to the dynamical equations. Because the moisture and vertical velocity fields both spin up in these methods, good initial precipitation rates can be obtained (Miyakoda et al., 1978; Fiorino and Warner, 1981).

The first group of dynamic methods involve forward and backward time integrations cycling around the current observational time with data inserted repeatedly until convergence is obtained (Nitta and Hovermale, 1969; Miyakoda and Moyer, 1968; Hayden, 1973; Temperton, 1976; Morel et al., 1971; Miyakoda et al., 1978). Methods of this sort were analyzed by Talagrand (1981b). When the model contains irreversible processes, principally precipitation, the forward-backward integration is not well defined (Miyakoda et al., 1978).

The second group, the so-called dynamic relaxation methods, have been used extensivly by the UKMO and others (Hoke and Anthes, 1976; Davies and Turner, 1977; Fiorino and Warner, 1981; Lyne et al., 1982; Bell, 1983). In these methods a term forcing the model state toward the observations is added to the governing equations. Special gravity wave damping procedures are required.

The current four-dimensional data assimilation scheme at GFDL is a compromise between the two above dynamic methods (Miyakoda et al., 1976; Miyakoda et al., 1978; Stern and Ploshay, 1983). The current model state at the main synoptic times (0000 GMT or 1200 GMT) is used as first guess fields for OI analyses of data in 2 h time windows centered at 0200, 0400, ... GMT. As the model advances in time at every time step the model state is replaced by a weighted average of the model state and the OI analysis corresponding to the time nearest the model time. The weight for the OI is determined by the reliability of the data used. Additional gravity wave balancing is required. Originally Euler backwards time stepping was used (Miyakoda et al., 1976); at present an NL Machenhauer NMI is used (Stern and Ploshay, 1983).

Since the effect of latent heat release has been shown to be so important in forecasting intense storms, it is natural to consider using observed precipitation rates to force the model by an externally specified heating function during an assimilation period. The technique of specifying latent heating is similar in some respects, and could easily be incorporated into a dynamic initialization assimilation method (Section 5.4). Precipitation rates are routinely measured by radar for small areas and are potentially available on a routine basis from microwave instruments: for example, Adler and Rodgers (1977) obtained rainfall rates from the NIMBUS 5 ESMR. In the two forecast impact studies to date, however, indirect indications were used to estimate rainfall rates. Molinari (1982a) estimated rainfall rates for Hurricane Anita (1977) from IR imagery calibrated against ground-based radar. Fiorino and Warner (1981) estimated rainfall rates for Hurricane Eloise (1975) from satellite imagery by assigning a rainfall rate, obtained from Adler and Rodgers (1977), to different categories of cloud systems, as defined by Gray et al. (1975). The vertical distribution of the heating had to be specified in these studies. Numerical modeling results were used as a guidance, but it is clear from the work reviewed in Section 3.6 that there is great sensitivity to the vertical distribution of heating. Both Molinari (1982a) and Fiorino and Warner (1981) obtained improved forecasts of intensity (relative to control experiments) when specified latent heating was used during a 12 h assimilation prior to the forecast. In addition, Molinari obtained improved track forecasts and Fiorino and Warner obtained improved precipitation forecasts. At the start of the forecast period there is a changeover from specified to model-generated heating; Molinari (1982b) has suggested a new closure for the parameter $\underline{b}$ in the Kuo (1974) convection parameterization, which would avoid this problem and allow one to use precipitation rates to specify $\underline{b}$ during an assimilation.

5.5 The Kalman Filter

The extended Kalman-Bucy (KB) filter yields in principle the optimum combination of all present and past data and thus solves the four-dimensional assimilation problem. All other approaches, including those discussed above, are therefore suboptimal. Here "optimum combination" means best of all linear combinations as measured by all quadratic loss functions. From time to time this solution has been suggested in the meteorological literature (Jones, 1965; Petersen, 1968; Talagrand and Miyakoda, 1971). However, the KB filter has never been pursued beyond some simple theoretical exercises, because unfortunately, its arithmetical requirements make it impractical.

Recently Ghil and coworkers have examined the KB filter in some idealized settings not so much as a prototype for an operational application but for the analysis of practical yet suboptimal systems such as OI (Ghil et al., 1981; Ghil et al., 1982; Cohn et al., 1981; Cohn, 1982). They have shown that the KB filter must be modified to suppress gravity waves. This so-called projected KB filter is optimum only with respect to a particular quadratic loss function (Cohn, 1982). It is found that the optimal correlation functions are not isotropic, especially at the upwind edge of a data rich area (Cohn et al., 1981). In the KB formalism the system noise and observational noise covariance matrices are presumed to be known. These matrices characterize the forecast and data errors respectively. Adaptive estimation of these covariance matrices is possible if some simplifying assumptions are made (Ghil et al., 1982).

5.6 Observing System Experiments, Data Rejection, Data Redundancy

The early observing system simulation experiments (OSSEs), which started with the Charney et al. (1969) experiment, led to real data observing systems experiments (OSEs) based on data from the NASA sponsored Data System Tests (DSTs) and more recently based on the FGGE data. Most of these experiments have been of the SAT versus NOSAT variety. In these studies two parallel four-dimensional assimilation experiments and associated forecast suites are run with and without satellite temperature sounding profiles. All other factors are held constant and the results are analyzed objectively by comparing skill scores and subjectively by comparing forecast maps.

The early OSE results, up to and including DST-6, were controversal. These results are summarized by Ohring (1979); generally speaking, for individual cases, the satellite temperature sounding profiles produced both positive and negative impacts and overall there was a small but positive average impact. The greatest impact was found in the Southern Hemisphere (Kelley et al., 1978), where there is the least conventional data and the greatest room for improvement. Studies which found small positive impacts include Atkins and Jones (1975), Bonner et al. (1976), Druyan et al. (1978), Halem et al. (1978) and Ghil et al. (1979). Similar results have recently been reported by Thomasell et al. (1983) and Druyan et al. (1983). However, the NMC group found no impact (Tracton and McPherson, 1977; Miller and Hayden, 1978; Tracton et al., 1980). Tracton et al. (1981) suggested that the NMC system was more advanced than systems used in other experiments, and thus in the NMC system there was little room for improvement when satellite data was added. That is, they argued that the remote sounding data were <u>redundant</u> in the NMC system because it makes such good use of the conventional data. To this date, NMC uses no satellite data over land (Morone and Dey, 1983). This is at best a partial explanation.

Several factors must be considered when evaluating impact tests (Gilchrist, 1982): (1) The soundings used were obtained statistically yet the situations when these extra observations are expected to be important are not necessarily "normal" situations which made up the bulk of the data base used to develop the statistical retrievals. (2) The analysis systems had been developed and tuned for conventional data. In particular, the statistical properties of the satellite observations - low vertical resolution, low horizontal variance, biases - were not properly taken into account. Phillips (1976) analyzed this point in detail. It is true that over oceans these statistics are difficult to estimate; just as the small ocean radiosonde data base makes it hard to develop statistical retrievals, it also makes it hard to evaluate statistical retrievals. (3) The sample of cases examined was small and may not have been representative. In some cases, where large positive impact was observed, the impact was traced to differences between the SAT and NOSAT initial conditions (Atlas, 1979, 1982; Brodrick, 1983). Also predictability varies from case to case; some cases are relatively insensitive to small changes in initial conditions. (4) Different models will accept or reject a particular type of data to different degrees. No attempt was made to tune the NMC system to the satellite data. Impact depends on the model used. Atlas et al. (1982) found that satellite data produced a larger positive impact when used in a higher resolution model. (5) Quality control is necessary in any system; there may be a substantial impact on a particular case due to the inclusion or exclusion of a particular radiosonde report which is borderline, i.e. which deviates enough from the first guess to be close to the rejection criterion.

A great number of OSEs have been conducted with the FGGE data sets (Gilchrist, 1982). Usually these experiments used the full set of FGGE observing systems as a control and examined the impact of removing one or several components. In particular there have been experiments examining the impact of removing each of the following: all satellite data (NOSAT), CTWs, temperature soundings, conventional upper air data (NORAOB), North Atlantic weather ship and Greenland radiosonde data, aircraft data, and drifting buoys. There are no reports of FGGE experiments defining the impact of moisture variables. Detailed results for the NOSAT and NORAOB experiments are presented by Halem et al. (1982). The GLAS group performed a series of further detailed experiments for the case of 21 January 1979 (Baker et al., 1983). The current general conclusions (Gilchrist, 1982) are that (1) satellite sounding data is vital to defining the large scale structure of the atmosphere, and (2) the impact on forecast skill is positive but not very large as seen by conventional measures, i.e., skill scores for continental areas. It appears that satellite data are largely redundant over land and the northern hemisphere. For this reason the greatest impacts have been obtained over Australia (Halem et al., 1982). Also, over the short range, 24-48 h, the beneficial effect of satellite data on defining the structure of the atmosphere over the oceans is only partly advected over the standard continental verification regions. This discussion suggests the strategy of verifying against a set of southern hemisphere ocean radiosondes.

During the FGGE year, 301 automatic drifting buoys were deployed south of 20°S (Guymer and LeMarshall, 1981). The enhanced FGGE data base resulted in the immediate improvement of analyses and forecasts in the Southern Hemisphere (Leslie et al., 1981; Guymer and LeMarshall, 1981; Bourke et al., 1982; Zillman, 1983); the buoys were especially important for the sea level pressure analyses and forecasts (Guymer and LeMarshall, 1981; Bourke et al., 1982).

Nonconventional wind observations - CTWs, scatterometer, and aircraft observations - are single level data. The degree to which these data are allowed to influence other levels is problematical. Vertical correlation functions may be calculated from radiosonde data. These correlations vary significantly with season and place (Blackmon et al., 1979; and references therein). Without substantial spreading of the data to other levels, it will tend to be rejected by the model. Therefore on empirically derived correlation function may be inappropriate for a model with only a few layers. Despite this difficulty, nonconventional wind observing systems are found to be very important in tropical analyses and forecasts and to a lesser degree in the extratropical southern hemisphere (Gilchrist, 1982). In the extratropics, when satellite sounding temperature data are available, single level winds are largely redundant since the atmosphere is so nearly geostrophic. Because of this and because forecast skill is low in the tropics, wind measurements are presently considered less valuable than temperature soundings (Gilchrist, 1982).

The impact of aircraft winds have been studied by Lorenc (1982) and Baede et al. (1983). The impact of CTWs was studied by Kållberg et al. (1982). These experiments and others were discussed by Gilchrist (1982). Level assignment, as discussed in Section 4.5, is a particular problem when using CTW. Recent simulation experiments of a lidar wind profiling system by Halem (private communication) show that this new observing system should be very valuable.

The impact of SEASAT scatterometer winds was found to be beneficial in a simulation experiment (Cane et al., 1981) but real data experiments have shown only small impacts in the northern hemisphere and larger but essentially unverifiable impacts in the southern hemisphere (Yu and McPherson, 1981, 1983; Atlas et al., 1982, 1983; Aune and Warner, 1983). The SASS data is difficult to use for two reasons. First there is the wind ambiguity problem. Second like all single level data, a vertical correlation function must be specified and this is especially difficult because the scatterometer senses winds at the bottom of the boundary layer (Fiorino and Warner, 1981). The relationship between winds at the surface and in the free atmosphere above the boundary layer depends crucially on the state of the boundary layer, especially its stability. The boundary layer parameterization used in the model could in principle be used to infer winds at the top of the boundary layer from observed surface winds using observed or forecast sea surface temperature, boundary layer temperature, boundary layer humidity, etc. These problems as well as the problem of the incomplete coverage of the actual satellite data were not addressed by the Cane et al. simulation study.

5.7 Implications

There are several important outstanding problems in data assimilation which affect our proposed research. These problems may be considered either obstacles to the success of our project or research opportunities. We take the latter point of view. Basically the problem of four-dimensional assimilation is to use all available data to influence the model state properly. By "properly" we mean without generating shocks to the model, without rejection of data and without extraneous gravity waves. As we have seen, considerable progress has been made in understanding and solving thse problems with respect to mass and momentum. However, there remain several important areas where progress is required. On the other hand, little attention has been paid to the inclusion of moisture in four-dimensional assimilation cycles. Furthermore, moisture data has its own characteristics which require special consideration.

We will end this section by raising a series of largely unresolved issues.

1) The time evolution of the atmosphere should be taken into account when using synoptic data. If we consider this problem in isolation, an ideal solution would be to assimilate data continuously (Ghil et al., 1979). However, continuous assimilation is associated with large shocks, data rejection and generation of gravity waves. Morel and Talagrand (1974) suggest that in the absence of an initialization procedure the time between insertions should be as large as or larger than the gravity wave dissipation time scale. Gravity waves can be removed with simple unconstrained NL NMI but this tends to increase data rejection, especially in a continuous assimilation cycle because only small amounts of data would be used at any one time. The linear incremental NMI approach of Puri (1983a) or the iterative OI/NMI approach of Williamson and Daley (1982) used in conjunction with a continuous assimilation might provide a practical solution. Other approaches should be considered. Within an intermittant assimilation

scheme, it might be useful to simply advect asynoptic observations of quasi-conserved quantities (e.g., potential temperature or total vorticity) until the next analysis time. The advecting velocity would be that forecast by the model. Other interesting theoretical approaches (Hoffman, 1981; LeDimet et al., 1982; Phillips, 1982) do not seem practical at the present time.

2) Currently, there are difficulties when using NMI in the tropics, presumably because the physical processes, especially the effect of latent heat release, are important factors contributing to the balance in the tropics (Tribbia, 1981). The effect of poor initial conditions in the tropics may adversely affect extratropical forecasts at forecast times as short as 24 h (Somerville, 1980). Tribbia's (1983) recent suggestion for higher order operational NL NMI, which incorporates physical processes easily and thus helps avoid these problems, should be investigated further.

3) More consideration must be given to the choice of moisture variable to be analyzed. Further research is required to provide adequate statistics for OI.

4) Model forecasts appear to be sensitive to small scale features in the moisture analysis. The impact of humidity data should be greatly enhanced by high spatial resolution in the model and in the observations.

5) The initialization or balancing of the moisture field with the mass and momentum field needs study. This problem is potentially very important for the 0-12 h prediction of precipitation, cloudiness, intense cyclogenesis and hurricanes. The impact of enhanced moisture analyses is probably greatly affected by whether or not, and to what extent the initial moisture and vertical velocity fields are balanced.

6) There is considerable evidence that moisture data will be rejected. This undoubtedly depends on the balance between the initial moisture and vertical velocity fields. Data rejection probably also depends on the parameterizations of the model and may be affected by tuning these parameterizations.

7) The assimilation of observations of diagnostic variables within the model must be addressed. Any variable which is given by a combination of model prognostic variables may be considered a diagnostic variable. Examples are rainfall rates, precipitable water, and cloud imagery. Although both the VAM and OI may be applied to this sort of problem, the VAM seems to be more easily applied to such data.

6. SUMMARY AND CONCLUSIONS

Common large-scale numerical weather prediction (NWP) practice and research have relegated moist processes and variables to secondary roles. (By "moist variables" we mean all those quantities needed to describe the distribution of water substance and cloudiness.) Until recently this situation was reasonable; the effort expended on NWP was directed at problems which would give the greatest impact. Now, however, with a prelude of three decades' experiences and steady advances in NWP, the greatest opportunity to advance global NWP probably lies in improving the parameterizations of moist processes and the observation and assimilation of moist variables. There is accumulating evidence that improvements in the physical parameterizations have positive impact on short range as well as medium range forecasts even in global models, despite the fact that observed cloud fields are not used to specify the initial state. The more complete use of satellite data to test and tune the physical parameterizations as well as to provide more comprehensive initial states and boundary conditions promises greater understanding and better predictions of short range weather changes.

Modern NWP models are often unable to predict rapidly developing synoptic scale intense storms. Within this class we include both tropical hurricanes and extratropical "bombs" (Sanders and Gyakum, 1980). This defect is the most spectacular deficiency of current models: <u>The failure of the models at scales and forecast times for which they should be valid</u> is troubling and suggests some physical process or interaction between physical processes is missing from the models. The most likely cause is that the latent heating is not properly parameterized in current models, although it remains a possibility that model resolution is still inadequate, especially vertical model resolution, to predict these storms accurately. There is considerable evidence that the generation of energy by intense storms involves an interaction between large scale dynamics and small scale cumulus convection. Even mesoscale models are not always able to predict these phenomena accurately. Experiments with mesoscale models do show the importance of the initial specification of moisture and of the parameterizations of moist processes in the development and maintenance of intense storms, but the evidence provided by forecast impact studies is mixed: Quantities like precipitation during the first few hours of the forecast are very sensitive to the initial moisture field. Forecasts of cyclone development are sensitive to latent heat release and to its vertical distribution. These two points together suggest that quantities like the 24 h prediction of geopotential should also be sensitive to the initial moisture field. The available evidence, however, suggests the opposite. This is an apparent contradiction. Generally, better precipitation forecasts are obtained when the initial moisture field includes high resolution data and/or when the initial vertical velocity and moisture fields are balanced or adjusted to maintain the currently observed precipitation. Based on this evidence we hypothesize that three requirements must be satisfied simultaneously in order to obtain significant positive impacts from moisture data: (1) high resolution moisture analyses which are (2) in balance with the divergent component of the wind analyses and (3) a reasonably accurate parameterization of convective latent heat release, which must provide the correct vertical distribution of this heating.

Accurate prediction of latent heating is essential for predicting intense storms and is therefore of utmost importance for improving current global NWP models. Accurate prediction of latent heating requires accurate parameterizations of condensation within large-scale and deep convective clouds, and adequate moisture predictions. Prediction of moisture depends on the initial specification of moisture as well as the parameterizations of evaporation and precipitation. Evaporation is partly controlled by the energy balance of the surface which is affected by clouds blocking solar radiation, ice and snow cover, sea surface temperature, etc.

The prediction of clouds by modern NWP models is no better than the use of persistence as a forecast out to 48 h. Since other meteorological quantities are skillfully forecast in this range, the lack of skill at forecasting clouds is generally considered evidence that clouds do not materially affect the dynamics of the atmosphere. (The Air Force cloud forecasting system is an exception to this discussion because it is more skillful than persistence by roughly 6 h and because clouds do not feedback on the dry dynamic model which is used to predict winds.) However, clouds are important in many physical processes occurring in the atmosphere. These processes - latent heat release, radiative heating, turbulent mixing, etc. - act to force the atmospheric fields of mass and momentum. Thus, the conclusion that clouds do not affect the forecast at short range is debatable, even excepting the intense storms which we discussed above, and this conclusion becomes less and less tenable as the forecast range increases. Besides the feedback between clouds and dynamics, cloud forecasts are of great intrinsic value - obviously so for aviation interests but also for forecasts of minimum and maximum surface temperatures (the quantities which interest the general populace) because of the modulation of radiative surface heating by clouds.

There appears to be room for substantial improvement in cloud forecasting. Operational NWP models currently do not explicitly predict clouds; clouds are diagnosed from predicted model variables by empirical relationships. A simple, but typical, cloud forecast method fills all model grid boxes with clouds for which the forecast relative humidity exceeds a critical value, e.g., 85%. Operational cloud forecasting schemes are all variants of this basic method.

There are two approaches to improving cloud forecasts: The first approach is to improve the forecast of humidity. Since current cloud forecasts use the forecast humidity to specify clouds, improving humidity forecasts should improve cloud forecasts. The second approach is to improve the parameterization of cloud physics. Currently some of the physical processes controlling cloud evolution are absent from operational models, but some research and mesoscale models have more complete cloud physics, with improved parameterizations of convective and large-scale precipitation and cloudiness, microphysics, interactions with radiation, etc. Some models explicitly predict cloud water amount. This requires the initial specification of cloud water amount. These two approaches, of course, are not mutually exclusive; they are complementary.

Since the primary rationale for designing a parameterization is to account for the effect of a particular physical subgrid scale process on the resolved scales, physically important interactions between different subgrid scale processes are sometimes neglected in NWP models. There are several important physical processes in this category which affect cloud evolution, and which depend on the parameterization of clouds. First, there is the interaction between clouds and radiation. This interaction is handled crudely at present, but is in reality complex. For example, the following processes may be important: Optically thin clouds can be dissipated (or enhanced) by radiation. The emissivity of cirrus clouds is controlled by cloud crystal size and shape spectra and cloud water amount. A discussion of the dynamic significance of the radiative effects of clouds is given by Kreitzberg

(1976). Second, subgrid scale turbulence affects the mixing of the cloud with its environment (i.e., entrainment and detrainment) and in turn turbulence is generated by the release of cloud buoyancy. Third the flux of moisture at cloud base intimately connects low clouds and the planetary boundary layer.

Thus, the parameterizations of all moist processes in global NWP models should be improved. The means exist: Some GCMs used in climate simulations and some research mesoscale models have advanced parameterizations, including those involving moisture, which can be adapted for global NWP models. Extending these parameterizations to include cloud water amount as a prognostic variable and to improve the interactions between the different parameterized processes is desired.

Improved moisture analyses are needed. The initial specification of humidity is certainly important for short term predictions of cloudiness and precipitation, but its importance for predictions of the "dry" atmospheric variables, i.e. surface pressure, geopotential heights, winds, and for predictions of long term (24 h and longer) moist variables is not established. However, in the tropics latent heating is an essential part of the balance of the atmosphere and the "spin-up" time for moisture in the tropics is 48 h or more. Further, for tropical and extratropical cyclones latent heating is an important source of energy. These facts suggest that initial moisture data should have a beneficial impact on large scale numerical weather predictions. Over oceans, and for scales not resolved by the radiosonde network, satellite observations of moisture variables are important. This is especially so for intense oceanic storms because the conventional observing platforms, ships and planes, actively avoid these storms. There are many observations of moisture potentially available from satellites which are not commonly used in global NWP models. For example, integrated liquid water from microwave observations, radiances from infrared channels sensitive to water vapor and cloud topography from visible and infrared channels all contain information about water variables. Since satellite data is incomplete and asynchronous, four-dimensional assimilation methods must be used. These methods should be specialized for moisture variables. In particular, better statistics and initialization procedures for moisture variables are needed.

Improved boundary condition data sets - sea surface temperature, sea ice extent, surface albedo, etc. - are also needed. These data are available or are potentially available from satellite observations.

SUMMARY AND CONCLUSIONS

High spatial resolution is vital when moisture is involved because it varies more, and on smaller spatial scales than other meteorological variables. The resolution currently attained by satellite infrared sounders, O(40 km), is adequate for global NWP models. Resolution in the forecast model and assimilation system is limited by computer resources but reasonably fine spatial resolution O(100 km) is attainable now for research purposes and should be feasible shortly on an operational basis.

Benefits which we expect to follow from the improvements discussed above include the following: (1) Improved forecasts of intense storms; (2) better precipitation and cloud forecasts (these are, after all, the weather items of most direct usefulness); (3) better initial and boundary conditions for mesoscale models (this may be the means of improving intense storm forecasts; a global data assimilation system is needed to provide initial conditions in oceanic areas); (4) better long-range forecasts, since improvements in short-range forecasting will also be mirrored at longer range (Lorenz, 1982). Since all models drift towards their own climatology and since this bias is a substantial part of the long-range forecast error, more realistic parameterizations should lead to more realistic model behavior at all scales.

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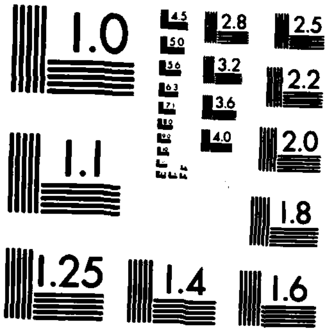

MICROCOPY RESOLUTION TEST CHART
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APPENDIX A

Variables used without definition in the text are defined here.

δ, Δ	difference, increment
F	cloud cover
g	gravitational constant
P_s	surface pressure
q	specific humidity; mixing ratio
r	water vapor mixing ratio
RH	relative humidity
t	time
T	temperature
$\underset{\sim}{V}$	horizontal velocity vector
z_α	height of constant pressure α
Z	height

APPENDIX B

Acronyms used in the text and references are defined here. For organizations listed below, parent organization in parentheses and location or current name in square brackets follow the acronym definition.

AFCRL	Air Force Cambridge Research Laboratory [now AFGL]
AFGL	Air Force Geophysics Laboratory [Hanscom AFB, MA 01731]
AFGWC	Air Force Global Weather Center (AWS) [Offut AFB, NE]
AMMS	Advanced Microwave Moisture Sounder
AMS	American Meteorological Society [Boston, MA 02108]
AMTS	Advanced Moisture and Temperature Sounder
ANMRC	Australian Numerical Meteorology Research Center [Melbourne, Victoria 3001, Australia]
AVHRR	Advanced Very High Resolution Radiometer
CP	Conference Publication
CPS	condensation pressure spread
CR	Contractor Report
CTW	cloud track wind
DMSP	Defense Meteorological Satellite Program
DOT	Department of Transportation [Washington, D.C.]
DST	Data Systems Test
EBBT	equivalent blackbody temperature
ECMWF	European Centre for Medium Range Weather Forecasts [Shinfield Park, Reading, Berkshire RG2 9AX, England]
EDIS	Environmental Data Information Service
ERL	Environmental Research Laboratories (NOAA)
ESMR	Electronically Scanning Microwave Radiometer
ESSA	[now NOAA]
GFDL	Geophysical Fluid Dynamics Laboratory (NOAA) [Princeton University, Princeton, NJ 08540]
FGGE	First GARP Global Experiment
GARP	Global Atmospheric Research Program
GATE	GARP Atlantic Tropical Experiment
GCM	global circulation model
GISS	GSFC Institute for Space Studies (NASA) [New York, NY 10025]

GLAS	GSFC Laboratory for Atmospheric Sciences (NASA) [Greenbelt, MD 20771]
GMT	Greenwich mean time
GOES	Geostationary Operational Environmental Satellite
GPO	Government Printing Office [Washington, D.C. 20402]
GSFC	Goddard Space Flight Center (NASA) [Greenbelt, MD 20771]
GTS	Global Telecommunication System
HIRS	High Resolution Infrared Sounder
IAMAP	International Association of Meteorology and Atmospheric Physics
ICSU	International Council of Scientific Unions
IP	Internal Publication
IR	infrared
JPL	Jet Propulsion Laboratory (NASA) [California Institute of Technology, Pasadena, CA 91103]
KB	Kalman-Bucy
LFM	Limited-area Fine Mesh
McIDAS	Man Computer Interactive Data Access System
MOS	model output statistics
MSU	Microwave Sounder Unit
NASA	National Aeronautics and Space Administration [Washington, D.C. 20590]
NCAR	National Center for Atmospheric Research [Boulder, CO 80307]
NCC	National Climate Center (NOAA) [Washington, D.C.]
NEMS	NIMBUS E Microwave Spectrometer
NEPRF	Naval Environmental Prediction Research Facility [Monterey, CA]
NESC	National Environmental Satellite Center [now NESS]
NESDIS	National Environmental Satellite, Data, and Information Service (NOAA) [Washington, D.C. 20233]
NESS	National Environmental Satellite Service (NOAA)
NL	nonlinear
NMC	National Meteorological Center (NOAA NWS) [Washington, D.C. 20233]
NMI	normal mode initialization
NOAA	National Oceanic and Atmospheric Administration (U.S. Department of Commerce) [Washington, D.C.]
NOO	Naval Oceangraphic Office [Washington, D.C.]
NTIS	National Technical Information Service [Springfield, VA 22161]
NWP	numerical weather prediction

NWS	National Weather Service (NOAA) [Silver Spring, MD]
OI	optimal interpolation
OLS	Operational Line-scan System
OSE	observing systems experiment
OSSE	observing systems simulation experiment
PBL	planetary boundary layer
PE	primitive equation
RP	Reference Publication
SASS	SEASAT-A scatterometer system
SCAMS	scanning microwave spectrometer
SCM	successive correction method
SDSD	Satellite Data Services Division (NOAA NCC) [Washington, D.C. 20233]
SIO	Scripps Institution of Oceanography [La Jolla, CA]
SIRS	Satellite Infrared Spectrometer
SMMR	Scanning Multichannel Microwave Radiometer
SMS	Synchronous Meteorological Satellite
SP	Symposium Publication; Special Publication
SSE	Special Sensor E
SSH	Special Sensor H
SSM	Special Sensor Microwave
SST	sea surface temperature
SSU	Stratospheric Sounder Unit
TM	Technical Memorandum
TN	Technical Note
TOVS	TIROS Operational Vertical Sounder
TR	Technical Report
TSC	Transportation Systems Center (DOT) [Cambridge, MA]
UKMO	United Kingdom Meteorological Office [Bracknell, UK]
UNESCO	United Nations Educational Scientific and Cultural Organization
VAM	variational analysis method
VAS	VISSR Atmospheric Sounder
VISSR	Visible and Infrared Spin Scan Radiometer
VTPR	Vertical Temperature Profile Radiometer
WB	Weather Bureau [now NWS]
WCP	World Climate Program

WMO	World Meteorological Organization [Geneva]
WWW	World Weather Watch
3DNEPH	three-dimensional nephanalysis
4-D	four-dimensional

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